Antarctic climate history and global climate changes

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Abstract

Antarctic climate changes have been reconstructed from ice and sediment cores and numerical models (which also predict future changes). Major ice sheets first appeared 34 million years ago (Ma) and fluctuated throughout the Oligocene, with an overall cooling trend. Ice volume more than doubled at the Oligocene-Miocene boundary. Fluctuating Miocene temperatures peaked at 17–14 Ma, followed by dramatic cooling. Cooling continued through the Pliocene and Pleistocene, with another major glacial expansion at 3–2 Ma. Several interacting drivers control Antarctic climate. On timescales of 10,000–100,000 years, insolation varies with orbital cycles, causing periodic climate variations. Opening of Southern Ocean gateways produced a circumpolar current that thermally isolated Antarctica. Declining atmospheric CO_2 triggered Cenozoic glaciation. Antarctic glaciations affect global climate by lowering sea level, intensifying atmospheric circulation, and increasing planetary albedo. Ice sheets interact with ocean water, forming water masses that play a key role in global ocean circulation.

Subjects: Earth Science, Climatology, Geology, Hydrology, Oceanography, Atmospheric Sciences
Keywords: Antarctica, climate, palaeoclimate, glaciation, cryosphere, climate change, ice sheets, palaeoenvironments.

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1 Introduction

In the 180 million years since the Gondwana supercontinent began to break up, its central fragment, Antarctica, has changed from a warm and equable environment with abundant plant and animal life to its current state, covered almost entirely by ice sheets totaling 27 million km^3 in volume (Fretwell et al., 2013). Despite its geographic isolation, Antarctica plays a major role in the global climate system, producing long-ranging influences through ocean currents, the atmosphere, and sea level: if the present-day Antarctic ice sheets melted, global sea levels would rise 58 m. There is thus good reason to study the past history of the Antarctic climate and to use this information to predict its future.

We begin this chapter with a brief tectonic history of Antarctica, charting its movements and those of its neighboring continents since the Gondwana breakup. We then introduce the main tools used in investigating Antarctic (and global) climate history: sedimentary and ice core archives and numerical modeling. The subsequent sections detail the climatic evolution of Antarctica since its formation and the underlying drivers that have shaped that evolution. We then continue with a history of the interactions between the Antarctic and global climates, leading into an overview of likely Antarctic changes in the next century and their global repercussions. Finally, we summarize the current state of Antarctic science and likely future research directions.

2 The Antarctic continent in the framework of the breakup of Gondwana

The Antarctic continent (figure 1) is centered asymmetrically around the South Pole, largely to the south of the Antarctic Circle (66°34′ S). It has an area of more than 14 million km^2 (about half the size of the United States), over 98% of which is covered by the Antarctic ice sheet, with an average thickness of around 2 km. At its thickest point the ice sheet is over 4.8 km deep.

Two hundred million years ago, Antarctica was the center of the Gondwana supercontinent (consisting of what are now Antarctica, India, Australia, South America, and Africa) and was not the dry and frozen continent we know today. About 180 million years ago, the movement of lithospheric plates, resulting in a hot megaplume that led to the formation of a triple junction, caused Gondwana to begin to break apart (Storey & Kyle, 1999). The rifting phase began in the Weddell Sea region in the Late Jurassic, during which East Gondwana (Antarctica, Australia, India, and New Zealand) and West

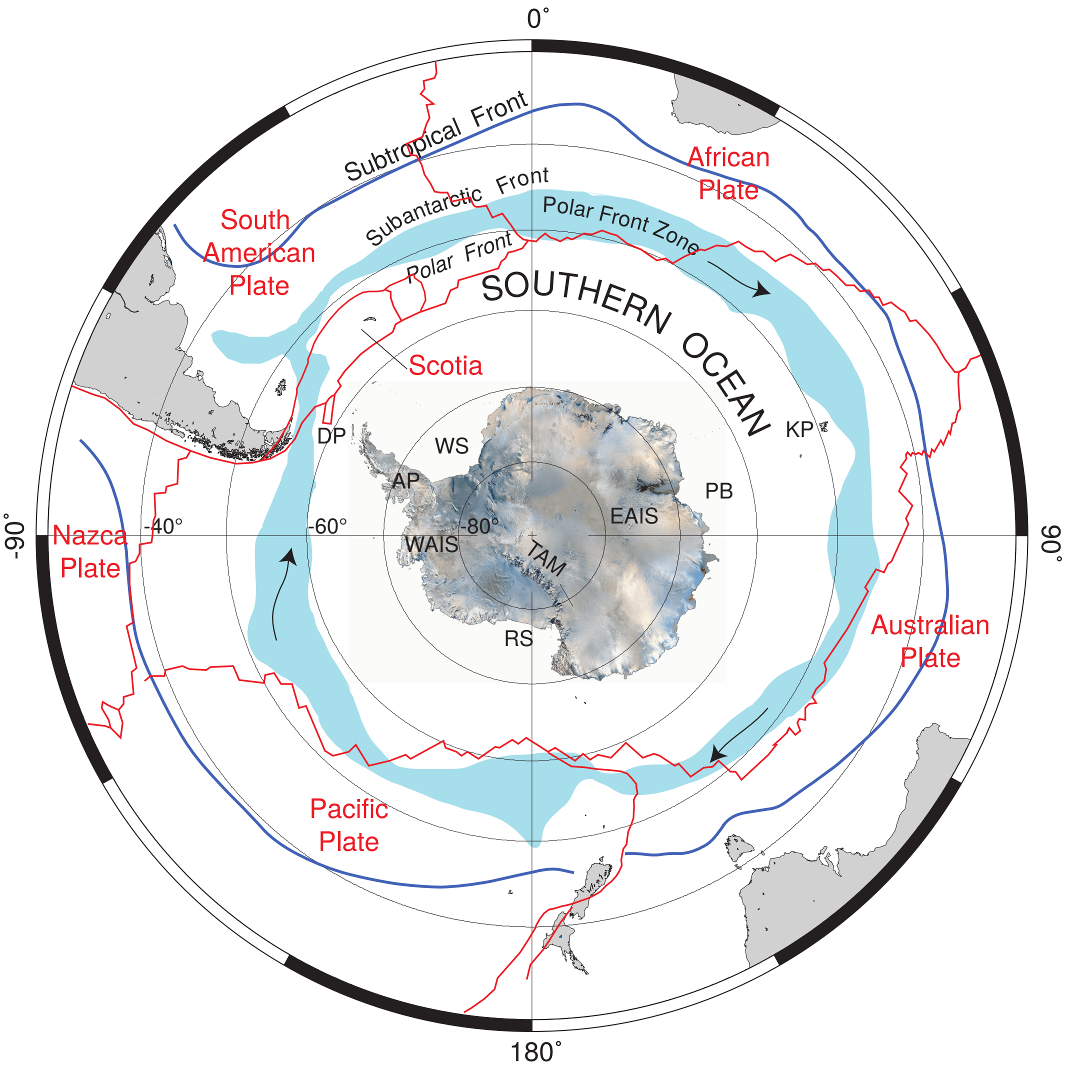

Figure 1: Antarctica, shown in polar stereographic projection to 30° S. Surrounding plate boundaries are outlined in red, and major oceanographic boundaries in blue. AP, Antarctic Peninsula; DP, Drake Passage; EAIS, East Antarctic Ice Sheet; KP, Kerguelen Plateau; PB, Prydz Bay; RS, Ross Sea; TAM, Transantarctic Mountains; WAIS, West Antarctic Ice Sheet; WS, Weddell Sea.

Gondwana (South America and Africa) moved apart (Lawver et al., 1992). The rifting proceeded in a clockwise direction around the Antarctic continent: India and Antarctica started to separate in the Early Cretaceous (Lawver et al., 1991), Australia and Antarctica started to separate in the Late Cretaceous (Veevers et al., 1990), and the Campbell Plateau, to the south of the New Zealand continent, rifted from Marie Byrd Land at 84 million years ago (Ma) (Lawver et al., 1991). During these tectonic phases, the Antarctic plate progressively moved toward southern polar latitudes, reaching its final location and configuration by the Late Cretaceous (DiVenere et al., 1994).

The polar position of the Antarctic Plate and the dispersal of plates and microplates eventually led to the development of a vigorous Antarctic Circumpolar Current (ACC) (e.g., Kennett 1978; Lawver et al., 1992; Barker et al., 2007). This set the scene for the establishment of continental glaciation, triggered by declining atmospheric pCO_2 levels (e.g., DeConto & Pollard 2003b; Royer, 2006; Peters et al., 2010). Glaciation affected the global climate, sea levels, ocean circulation, and atmospheric composition and dynamics and led to the present-day cold polar climate. The final barriers to circumpolar flow were to the south of Tasmania and of South America (the Drake Passage), with the deep-water opening of the Tasmanian channel constrained to before 32 Ma. Estimates of the time of opening of the Drake Passage have varied widely, from 41 Ma to 6 Ma (Barker et al., 2007).

3 Records of Antarctic climate

For the most recent 800,000 years (800 kyr), Antarctic ice cores provide valuable records of both regional and global climate signals (Lüthi et al., 2008). A particularly powerful feature of ice core archives is their incorporation of air samples in tiny bubbles, which allows the composition of ancient atmospheres to be reconstructed without resorting to indirect proxies. Antarctic ice core records could potentially be extended as far back as 1.5 Ma (Fischer et al., 2013). Beyond that time range, ice becomes more of a hindrance than a help to climate reconstruction: sediment successions around the Antarctic margin can be complicated or erased by cycles of glacial advance and retreat, and the present-day ice sheets restrict access to sedimentary archives beneath them. Furthermore, ice margin records are often poor in the carbonate microfossils widely used for age determination and paleoclimatic reconstruction, forcing increased reliance on techniques such as magnetostratigraphy and dinoflagellate analyses (e.g., G. S. Wilson et al., 1998).

Despite these challenges, many valuable sedimentary records have been cored since the 1970s from onshore sites, through ice shelves, and in the open waters around the Antarctic margin (Barrett, 2008). The first direct record of Oligocene Antarctic glaciation was retrieved by Deep Sea Drilling Project (DSDP) Leg 28 in 1973 from the eastern Ross Sea; it included glacial deposits up to 25 million years (Myr) old (Hayes et al., 1975). In 1973–5, the Dry Valley Drilling Project conducted the first onshore Antarctic scientific drilling (as well as one experimental coring from a sea ice platform), recovering sediments of up to Miocene age from the McMurdo Dry Valleys; those records provided a history of glacial-interglacial cycles (Webb & Wrenn, 1982). In 1979 the McMurdo Sound Sediment and Tectonic Studies Project drilled through sea ice to a depth of 227 m, reaching Oligocene sediments and providing further evidence of cyclic glacial advance and retreat (Barrett et al., 1987).

From the 1980s onward, with the technical expertise gained from earlier projects, efforts to core sediment through sea ice became more ambitious and more successful. In 1986 the CIROS-1 core was drilled 702 m below the sea floor under McMurdo Sound ice in the Ross Sea; it was the first to extend as far back as the Eocene and thus the first to record the inception of Antarctic glaciation at the pivotal Eocene-Oligocene transition (Hambrey et al., 1989). In the 1990s three sites cored in McMurdo Sound by the Cape Roberts Project (CRP) (Davey et al., 2001) provided the first evidence of the Antarctic response to orbital forcing in the Oligocene and Miocene (Naish et al., 2001). The CRP was succeeded by the ANDRILL McMurdo Ice Shelf (MIS) (Naish et al., 2007, 2008) and Southern McMurdo Sound (Harwood et al., 2009) projects in 2006–7, which drilled cores extending from the Oligocene into the Pleistocene, temporally overlapping ice core records at the younger end and CRP cores at the older end. Notable results from ANDRILL include the first Antarctic record of the Mid-Miocene Climatic Optimum (MMCO) and evidence of open-water conditions in the Ross Embayment during the Pliocene.

The onshore and ice shelf drilling projects in the Ross Sea and McMurdo Sound regions have been complemented by ocean drilling in other areas of the Antarctic continental shelf: Ocean Drilling Program (ODP) Legs 119 (Barron et al., 1989) and 188 (O'Brien et al., 2001) in Prydz Bay (eastern Antarctica); ODP Leg 178 (Barker et al., 1999b) and SHALDRIL (Wellner et al., 2011) around the Antarctic Peninsula; and Integrated Ocean Drilling Program (IODP) Leg 318 at the Wilkes Land margin (Escutia et al., 2011). These expeditions have produced sedimentary records covering the entire Oligocene-to-Pleistocene time span, paralleling the McMurdo Sound record from CRP and ANDRILL.

More-distal sedimentary archives provide more-extensive and more-complete records, although they reflect less direct influence of the Antarctic climate. The most widely studied climate signal from such records is the ratio of two oxygen isotopes, ^{16}O and ^{18}O, in carbonate, commonly abbreviated $\delta^{18}O$. Interpretation of the $\delta^{18}O$ value is not always straightforward, since it is affected by both water temperature and global ice volume. To reliably infer ice volume from a $\delta^{18}O$ value, one can use other methods to constrain the water temperature. Frequently used temperature proxies include the foraminiferal magnesium/calcium ratio (Rosenthal et al., 1997) and the TEX_{86} proxy, based on the compositions of planktic lipids (e.g., Schouten et al., 2002). Several other techniques have also been employed to resolve the ambiguity of $\delta^{18}O$ records (Shakun et al., 2015).

Another widely available proxy for glaciation is the change in global sea level (eustasy), which can be inferred, for example, from backstripped stratigraphy of suitable coastal areas (e.g., Miller et al., 2005): when more water is sequestered on land as an ice sheet, the sea level drops correspondingly. While eustasy is affected by several other factors, ice sheet growth and decay is the only one capable of producing fluctuations of more than 10 m on timescales of less than 10 kyr (Miller et al., 2005). Thus when sea level changes are sufficiently large and rapid, a glacial influence can be inferred.

Neither the glacial component of a $\delta^{18}O$ record nor a global sea level record can say anything explicit about where the ice was located, and more direct sedimentological evidence or modeling informed by known paleogeography is needed to apportion the inferred ice cover between Antarctica and other potential host areas.

4 Modeling of Antarctic climates

Numerical models are a vital part of climate investigations, not only for predicting future developments but for constraining the mechanisms and causes of climate change recorded in physical archives.

Geological records, when correlated by thorough chronostratigraphy, provide vital information on the mechanisms of climate change, but records alone are insufficient to definitively distinguish causal relationships from chance correlation. Here, modeling-based studies provide a vital link, allowing hypotheses to be developed and tested against the available paleo-records. In Antarctic paleoclimatology, one of the essential tasks of climate models has been determining the most important triggers and drivers for the onset of glaciation. For instance, the modeling study of DeConto and Pollard (2003b) demonstrated that despite the temporal correlation between the opening of ocean gateways and the onset of major Antarctic glaciation in the Oligocene, declining CO_2 would have been sufficient to initiate glaciation even without the presence of gateways.

Considering the inevitably partial nature of paleoclimatic records, modeling also plays an important role in reconstructing events not directly recorded in geological archives. For example, Barker et al. (1999a) used a numerical ice sheet model, constrained by sedimentary records around the Antarctic margin, to infer a history of ice sheet growth and decay. Modeling also plays an important role in planning future field research: model studies can determine likely locations for sediments or ice cores of a certain age or type, allowing sampling to be more precisely targeted to answer particular questions or fill known gaps in existing data sets.

The first general circulation climate models were developed in the 1950s (Lynch, 2008). Since then, better understanding of the physical processes of climate, along with continuing, exponential increases in computing power, has allowed climate models to improve vastly in power and sophistication. Numerical ice sheet models have followed a similar trajectory. In early paleoclimatic models, ice sheets were often simulated using highly simplified, two-dimensional representations (e.g., Weertman, 1976). The development of more-sophisticated, three-dimensional ice sheet models (e.g., Mahaffy, 1976) soon led to their integration into model-based paleoclimate investigations (Oerlemans, 1982), and it is now possible to fully couple three-dimensional ice sheet models with ocean and atmosphere models in order to simulate the evolution of the entire system over long (millions of years) periods (e.g., DeConto & Pollard, 2003a). The physical sophistication of ice sheet models continues to improve, with recent developments including improved simulation of ice cliffs and ice shelf pinning effects from bathymetric rises under the ice (Pollard et al., 2015).

5 Antarctic climate evolution

Antarctica has occupied polar latitudes since the Early Cretaceous (Scotese, 2001) but has not been extensively glaciated for most of its history. Evidence from fossil vegetation indicates that Antarctic glaciation was rare or absent during the Cretaceous (Spicer & Corfield, 1992). Dramatic fluctuations in reconstructed sea levels (e.g., Miller et al., 2005; Galeotti et al., 2009) suggest that transient glaciations may have taken place even in the Cretaceous greenhouse world, but the sea level data are difficult to reconcile with the lack of evidence of glaciation in the $\delta^{18}O$ record (Ando et al., 2009; Miller, 2009).

Throughout the Cenozoic era, the overall history of Antarctic climate is one of increasing cooling and glaciation, but the growth of the ice sheets did not proceed in a steady, monotonic fashion, and the long-term cooling trend has been punctuated by a number of shorter temperature excursions in both directions (figure 2). There is no direct evidence of pre-Oligocene Cenozoic Antarctic continental glaciation, though Hollis et al. (2014) considered reduced temperature and lower sea levels in the Early Paleocene in the southwest Pacific

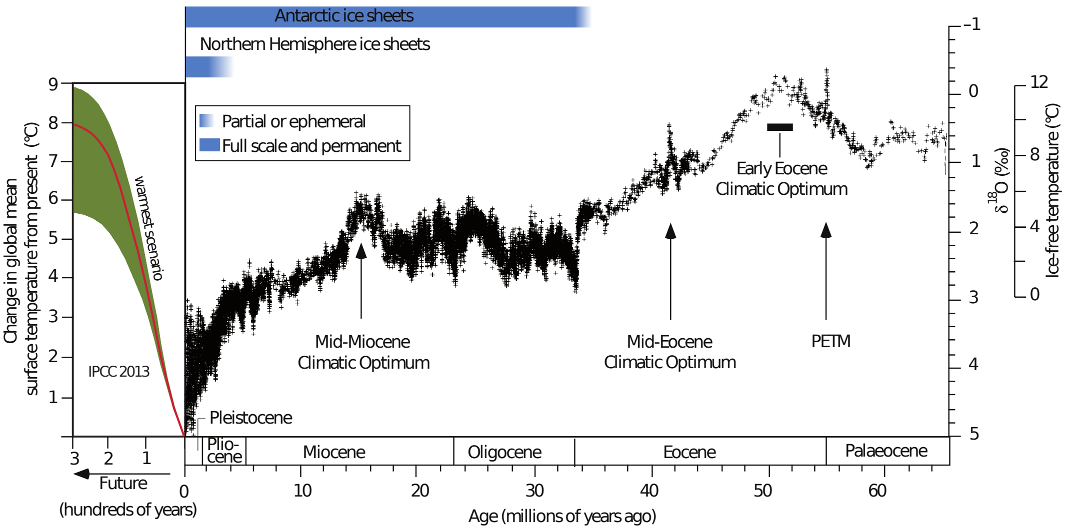

Figure 2: Global climate over the past 65 million years. The data are stacked from deep-sea oxygen isotope records based on data compiled from DSDP and ODP sites. The temperature scale on the right axis was computed on the assumption of an ice-free ocean and therefore applies only to the time preceding the onset of large-scale glaciation on Antarctica (isotope data from Zachos et al., 2008). On the left, global annual mean surface temperature projections for the future show the increase in temperature under an "unrestricted" scenario (IPCC RCP8.5) with continuing greenhouse gas emission beyond 2100 (Barrett, 2003). PETM, Paleocene-Eocene Thermal Maximum. Modified from Florindo and Pekar (2015).

to indicate transient ice sheets on Antarctica. The inception of continent-wide ice sheets has long been recognized as occurring in the earliest Oligocene (Wise et al., 1991; Zachos et al., 1992) on the basis of glacial deposits near shore and ice-rafted debris offshore. The late Paleocene and Eocene, which included the time of peak Cenozoic warmth (the Mid-Eocene Climatic Optimum), were punctuated by brief, extreme global warming events termed hyperthermals (Lourens et al., 2005), so significant Antarctic glaciation was unlikely.

The first major, well-recorded expansion of Cenozoic glaciation came at the start of the Oligocene (~34 Ma): global marine sediments record a sudden increase of more than 1‰ in $\delta^{18}O$, termed the Oi-1 event (Miller et al., 1991). Subsequent analysis of oxygen isotope and sea level records (Miller et al., 2009) has refined the date of Oi-1 to 33.55 Ma and identified two smaller precursor events, EOT-1 (33.80 Ma) and EOT-2 (33.63 Ma); it thus appears that the transition from the warmer Eocene world to the Oligocene icehouse occurred in three discrete steps. Records from around Antarctica itself indicate that this oxygen isotope shift was associated with a major expansion of the ice sheets, as recorded, for example, by ice-rafted debris and other sedimentological evidence from Maud Rise and the Kerguelen Plateau (Ehrmann & Mackensen, 1992) and by diamicts deposited in Prydz Bay (Hambrey et al., 1991). While most of the direct evidence for glaciation relates to East Antarctica, there are also indications that the northern Antarctic Peninsula was glaciated to sea level in the early Oligocene (Ivany et al., 2006), although the dating of this evidence has been called into question (Marenssi et al., 2010). Establishing the precise details of Oligocene glaciation has proved challenging because core records from the Antarctic margins are difficult to date precisely and often suffer from poor recovery of cored material and major depositional hiatuses. The dating of the CIROS-1 core, the first to capture the Eocene-Oligocene transition, has been revised repeatedly, but the current age model (G. S. Wilson et al., 1998; Roberts et al., 2003) confirms a cooling trend during the transition itself, following a relatively warm Eocene interval. The transition is also recorded in the CRP-3 core, which, although difficult to date precisely in this interval (Florindo et al., 2005), also seems to indicate sharp cooling at around the same time, as interpreted from a sharp decrease in magnetite concentration (Sagnotti et al., 2001).

For Antarctica, the Oligocene epoch was colder and more glaciated than the Eocene, but the ice sheets experienced repeated cycles of growth and decay, as indicated by large, rapid changes in global sea level (Kominz & Pekar, 2001) and $\delta^{18}O$ variations in deep-ocean sediment (Pälike et al., 2006). Oligocene glaciation recurred at periods corresponding to variation in the eccentricity and obliquity of the Earth's orbit, indicating that the climate system was responding to orbital cycles (Pälike et al., 2006). Despite the fluctuations, the overall Oligocene trend was toward colder conditions and increasing ice cover. Pollen and spore analyses from the CRP-2/2A core in the Ross Sea agree with this pattern, indicating a decrease in both abundance and variety of vegetation through the Oligocene (Askin & Raine, 2000). In contrast, decreasing $\delta^{18}O$ values in cores from beyond the Southern Ocean are suggestive of Antarctic deglaciation in the late Oligocene. However, that trend is absent in southern records, making it likely that the $\delta^{18}O$ decrease in more northerly basins records regional oceanic warming rather than Antarctic deglaciation (Pekar et al., 2006).

It is likely that the Oligocene also saw the initial establishment of an Antarctic Circumpolar Current (ACC), made possible by the continuing widening of the Drake Passage. While the timing of these events is still poorly constrained, most studies have placed the development of the ACC somewhere in the Oligocene (Barker et al., 2007). The ambiguity in dating the initiation of the ACC is caused in part by the fact that it may have developed gradually over an extended time: shallow ACC circulation may have preceded deep-water flow by millions of years (Lyle et al., 2007). Regardless of the exact timing, the initiation of the ACC was an important event for both Antarctic and world climate, isolating the Antarctic continent thermally and helping to form the stratified oceanic water structure that has persisted to the present day (Katz et al., 2011).

The Oligocene-Miocene boundary at ~23 Ma constitutes another pivotal moment in Antarctic climate evolution; in deep-ocean sedimentary records it is distinguished by a positive $\delta^{18}O$ excursion of around 1‰, termed the Mi-1 event (Miller et al., 1991). While the event itself lasted only around 200 kyr (Zachos et al., 2001), it was dramatic in scale, with ice sheets estimated to have grown from 50% to 125% of the size of the present-day East Antarctic Ice Sheet (EAIS) (Pekar et al., 2006). It was the first of a series of similar "Mi-" oxygen isotope events that, while less extreme than Mi-1, were implicated in the progressive cooling and glaciation experienced by Antarctica during the Miocene. As in the Oligocene, however, the cooling was not monotonic, and the periodic swings between cooler and warmer climate were driven by variation in the Earth's orbital parameters, as well as by changes in the oceanic carbon cycle (Zachos et al., 1997; Holbourn et al., 2005). The Mi-1 event itself coincides with an unusual set of orbital parameters: low orbital eccentricity and low-amplitude variance in obliquity, resulting in an extended period of cool summers that allowed ice sheets to accumulate (Zachos et al., 2001). After the initial onset of and recovery from Mi-1, the early Miocene experienced relatively warm intervals; palynological evidence from the Ross Sea indicates a regular recurrence of favorable growing conditions, reduced glacial activity and sea ice, and temperatures above freezing (Griener et al., 2015).

The warming cycles of the early Miocene culminated in the MMCO at around 17–14 Ma, immediately followed by the Middle Miocene Climate Transition (MMCT) at 14.2–13.8 Ma (Shevenell et al., 2004). The transition, as recorded in deep-ocean $\delta^{18}O$ variations, was a significant one: the MMCO was the warmest interval since the Eocene, while the MMCT, and continued cooling during the rest of the Miocene, produced temperatures colder than any previously seen in the Cenozoic. During this time, there was a sizeable EAIS, as inferred both from $\delta^{18}O$ and sea level curves (Barker et al., 1999) and from direct sedimentological evidence from the Antarctic continent itself (Lewis et al., 2007).

The general cooling trend of the late Miocene continued into the Pliocene and Pleistocene. A mid-Pliocene warm period (Salzmann et al., 2011) was succeeded by the third and last of the major Cenozoic global cooling events (the first two being the Oligocene-Miocene cooling and the MMCT). The Pliocene cooling event, usually dated at around 3–2 Ma, is chiefly associated with the onset of major Northern Hemisphere glaciation (Raymo et al., 1992) but has also been considered to mark the expansion of the EAIS to its modern extent, along with the development of its marginal ice shelves. The appearance of large ice masses in the Northern Hemisphere poses a challenge for reconstructions of Antarctic glaciation: from that point on, the global $\delta^{18}O$ record—which shows a regular sequence of glacial-interglacial cycles (Lisiecki & Raymo, 2005)—is affected not only by Antarctic ice mass and water temperature but also by Northern Hemisphere ice mass. For the West Antarctic Ice Sheet (WAIS), Antarctic sedimentary records (Naish et al., 2009) and modeling results (Pollard & DeConto, 2009) indicate dynamic behavior, largely as a result of its grounding below sea level. The EAIS, with its base on land, has been less susceptible to melting during interglacials, but there is no firm consensus as to its behavior during the Pliocene and early Pleistocene; there are some indications of long-term stability since 3 Ma (Denton et al., 1993) and other evidence suggesting more dynamic behavior until 1 Ma (Raymo et al., 2006). During the past million years, the EAIS has continued to respond to glacial-interglacial cycles, in a more limited fashion than the WAIS and northern ice sheets: sedimentological and modeling evidence indicate some thinning and retreat of the EAIS during the last glacial termination, but the extent of its response is minor in comparison with the WAIS (Mackintosh et al., 2011).

6 Drivers of Antarctic climate change

The complex and variable history of Antarctic climate has motivated much work on the causes of the climate changes. In recent years, an extra impetus for this research has appeared: anthropogenic emissions of carbon dioxide are pushing atmospheric CO_2 concentrations past 400 ppm, to levels not experienced since the Pliocene. It is therefore important to understand the role that CO_2 played in past warm Antarctic and global climates, in order to predict the effects of the current dramatic rise in its atmospheric concentration. Greenhouse gases are not the only control on Antarctic climate evolution: tectonics and variations in the Earth's orbit around the sun also play major roles. Establishing the relative importance of these drivers and understanding the complex ways in which they interact have been major themes in Antarctic research.

One of the fundamental drivers of global and Antarctic climate is incoming solar radiation. On timescales of tens to hundreds of thousands of years, the most significant variations in insolation are caused by regular variation in the shape of the Earth's orbit around the sun. These orbital variations, which affect not only the total amount of solar radiation reaching the Earth but also its distribution by latitude and by time of year, are usually referred to as Milankovitch cycles in honor of Milutin Milanković, who produced the first orbitally derived reconstructions of insolation cycles (Milankovitch, 1930). The three main orbital cycles arise from variations in precession of the equinoxes, with a period of ~21 kyr; obliquity of the ecliptic (axial tilt), with a period of ~41 kyr; and eccentricity of the orbital ellipse, with a period of ~100 kyr (Hays et al., 1976). The mechanisms by which these orbital parameters act on global climate are complex and still not fully understood, but many sedimentary archives record climate variations (most frequently in the form of $\delta^{18}O$ values) with periodicities close to the Milankovitch frequencies, providing strong evidence for their major role in driving climate change (e.g., Hays et al., 1976; Zachos et al., 1997; Warnaar et al., 2009). More-direct evidence for the orbital pacing of Antarctic ice sheets comes from the sequence stratigraphy of shallow marine cores from the Ross Sea (Naish et al., 2001).

Insolation variations are not synchronous across latitudes: when Northern Hemisphere summers are at their hottest, Southern Hemisphere summers are at their coldest. Despite this, glaciations occur synchronously across the globe, coinciding with cold Northern Hemisphere summers. Partly for this reason, Milankovitch cycles have traditionally been thought to control climate through their effect on glaciation in high northern latitudes, with the Antarctic response initiated indirectly through oceanic or atmospheric heat transport (e.g., Imbrie et al., 1992). Other explanations have been proposed (Huybers, 2009), including the idea that Antarctic ice is more sensitive to the duration than to the intensity of summer insolation (Huybers & Denton, 2008).

It has long been accepted that ocean gateways between Antarctica and adjoining continents have played a significant role in the development of glaciation. However, the extent and nature of this role are still debated. One obstacle to investigation has been the difficulty in establishing an exact chronology of initial gateway openings and subsequent expansions: since ocean currents are controlled by a gateway's size and shape, a full paleoceanographic reconstruction requires, in effect, a four-dimensional data set: a three-dimensional bathymetry developing through time. Despite the difficulties in reconstructing such data from the available records, the history of the circum-Antarctic gateways has been sufficiently well constrained for the development of theories linking it to glaciation. Kennett (1977) presented a model of Antarctic glacial development heavily influenced by ocean gateways, linking the onset of continent-wide glaciation in the mid-Oligocene to the development of a "relatively unrestricted circum-Antarctic current." That current in turn was made possible by the increasing distance between the Australian and Antarctic continents, and possibly by the initial opening of the Drake Passage between South America and Antarctica.

While earlier work on Antarctic climate frequently invoked gateways as the dominant driver for glaciation, more recent studies have placed increasing emphasis on the role of atmospheric CO_2 concentrations. A landmark modeling study by DeConto and Pollard (2003b) found that a steady decrease in atmospheric CO_2 from 1120 ppm to 560 ppm over 10 Myr would result in the growth of a continental-scale EAIS even with the Drake Passage closed. Subsequent work has been informed both by improved CO_2 reconstructions for the Cenozoic (e.g., Pagani et al., 2005) and by revised constraints on the timing of gateway openings (e.g., Livermore et al., 2007), but more-reliable and extensive data are still needed for a thorough understanding of these influences (Ruddiman, 2010).

Tectonics have influenced Antarctic climate not only through the distribution of land masses and ocean gateways but through the topography of the continent itself. Geomorphologists have argued that the present-day depressed East Antarctic interior and uplifted margin is a consequence of erosion rather than tectonism, resulting from the work of a dome-shaped ice sheet since its inception at around 34 Ma (Jamieson & Sugden, 2008). However, tectonism has had a profound effect on the Antarctic ice sheet through the post-Eocene formation of the West Antarctic Rift System. This resulted in crustal thinning that reduced the area of subaerial Antarctica by 20% (D. S. Wilson et al., 2012), thus making the marine-based WAIS inherently vulnerable (D. S. Wilson et al., 2013) in late Cenozoic times.

Results of modeling studies indicate that surface elevation can locally affect ice thickness and drainage patterns (Kerr & Huybrechts, 1999), but when compared with the effect of CO_2 concentration, elevation does not seem to be a major influence on the timing of the actual inception of ice sheets (DeConto & Pollard, 2003a).

Recent studies continue to argue for the importance of both tectonics (e.g., Bijl et al., 2013) and CO_2 (e.g., Pagani et al., 2011) as initiators of Antarctic glaciation, and modeling work (e.g., Sijp et al., 2009) continues to refine our understanding of the way in which those influences interact. There has also been much research focused on investigating the interactions among Milankovitch cycles, glaciations, and CO_2 variations. One long-standing focus of such research is the so-called 100-kyr problem (Imbrie et al., 1993). This term refers to the fact that for the past million years, climatic cycles have followed the 100-kyr cycle of orbital eccentricity; the "problem" is that compared with the precession and obliquity cycles, eccentricity does not produce a significant amount of solar forcing: the climate appears to be responding strongly to a weak orbital forcing and weakly to the stronger forcings. In the decades since Hays et al. (1976) provided strong confirmation of the 100-kyr climate cycle, dozens of models have been and continue to be proposed to explain it, often invoking ice and CO_2 feedbacks as part of an amplification mechanism. Theories include control by the carbon cycle via atmospheric CO_2 (Shackleton, 2000), synchronization of North American glacial cycles to eccentricity (Ganopolski & Calov, 2011), modulation of precession cycles by eccentricity (Ruddiman, 2003), the influence of orbital parameters induced by other planets (Berger et al., 2005), and synchronization of internal climate oscillations to a longer, 413-kyr eccentricity cycle (Rial et al., 2013). To date there is no consensus as to the exact mechanism behind the 100-kyr cycles.

A similar, but less extensively studied, problem exists for the climate cycles of the late Pliocene to early Pleistocene. During this time, global ice volume varied at a periodicity of 41 kyr, corresponding to the orbital obliquity cycle, rather than (as classical Milankovitch theory would predict) the 21-kyr precession cycle (Raymo & Nisancioglu, 2003). As with the 100-kyr problem, this contradiction has generated a number of proposed solutions (e.g., Huybers, 2006; Tabor et al., 2014), but as yet there is no strong consensus on the true explanation.

7 Antarctica and global climate change

Despite its isolated position, the Antarctic continent has a major influence on global climate and ocean systems. Variations in the volume of the ice sheets affect sea level by sequestering water on land: the current volumes of the East and West Antarctic Ice Sheets correspond to around 53 and 4 m of global sea level, respectively (Fretwell et al., 2013). In addition, the ice sheets are responsible for regional as well as global variations in sea level through direct gravitational attraction of water masses (Kuhn et al., 2010).

Where the Antarctic ice sheets reach the ocean, they exert a profound influence on the global organization of oceanic water masses. Antarctic Bottom Water forms at the Antarctic margin and constitutes one of the densest water masses on the planet (Orsi et al., 1999); this sinking of upper-layer water to abyssal depths is a vital link in the thermohaline circulation of the world ocean, which in turn affects climate worldwide. Within the same system, the upwelling of Circumpolar Deep Water around Antarctica is equally important: it provides a rare opportunity for the exchange of gases—notably, carbon dioxide—between these isolated deep waters and the atmosphere. The sharp rise in atmospheric CO_2 during the last deglaciation has been attributed to enhanced upwelling's allowing the ventilation of previously sequestered CO_2 from deep waters back into the atmosphere (Anderson et al., 2009). Antarctica also affects the world ocean system via the wind-driven ACC, the strongest ocean current on the planet (Barker & Thomas, 2004). The ACC, while helping to isolate the Antarctic continent thermally, also plays a key part in global oceanic circulation. As the only current connected to all the major oceans, it redistributes North Atlantic Deep Water from the Atlantic to other ocean basins (Rintoul et al., 2001).

Antarctic glaciation affects the atmosphere as well as the ocean. The strong temperature gradient between the glaciated polar region and warmer lower latitudes intensifies global atmospheric circulation (Flohn, 1984). Glaciation can also increase ocean productivity (and thus CO_2 drawdown) by fertilizing the ocean with an increased supply of iron-rich dust (Martin, 1990).

Another potential global-scale influence of Antarctic glaciation is exerted through the high albedo of ice and snow cover compared with the underlying land or water. A glaciated Antarctica reflects a higher proportion of incoming solar radiation, effectively changing the planet's radiation budget. Snow and ice albedo acts as a global positive feedback, since increased glaciation reduces absorbed radiation, thereby reducing temperature and supporting further glaciation; conversely, melting snow and ice cover increases absorbed radiation, increasing temperature and supporting further deglaciation. This amplification effect is not restricted to the glacial area itself: Hall (2004) found that surface albedo feedback amplified CO_2-induced warming by around 20% even in the tropics. Modeling has suggested that the effects of albedo variations in the Antarctic are less broad-ranging than those at lower latitudes (Ogura and Abe-Ouchi, 2001), but their influence nevertheless extends well beyond the continent itself.

As a result of these strong oceanic and atmospheric connections, the first major expansion of Antarctic glaciation at the start of the Oligocene was a pivotal change not only for the Antarctic continent but for the climate of the entire planet: meridional temperature gradients increased in the oceans; thermohaline circulation intensified; ocean acidity decreased and the calcite compensation depth increased; and katabatic winds caused increased upwelling in the Southern Ocean (Miller et al., 2009). The Oligocene-Miocene event was perhaps the most significant glacial episode for reconfiguration of the world ocean system, but every large fluctuation in the Antarctic ice sheets has wide-reaching effects, most obviously on global sea level.

8 Near-future changes in Antarctic climate and their global effects

The history of Antarctic climate reflects processes operating on a huge variety of timescales. Some of these, such as plate tectonics and the 400-kyr "long eccentricity" orbital cycle, act across time spans longer than the present tenure of *Homo sapiens sapiens* on Earth. At the other end of the range are mechanisms operating at century and subcentury scales, such as glacier calving and ice shelf collapse. In predicting future developments of Antarctic climate, there is naturally a strong focus on the decadal and century-scale processes that will directly affect our civilization. However, even a pragmatic, relatively short-term evaluation of future Antarctic climate and its global effects cannot afford to ignore longer-term processes, since the long-term and short-term processes continually affect each other.

On the one hand, rapid processes can trigger longer-term trends. For example, the current destabilization and retreat of glaciers in the Amundsen Sea sector of West Antarctica is occurring on timescales of decades to centuries (Joughin et al., 2014; Rignot et al., 2014), but model results suggest that their disappearance will result in a longer-term WAIS collapse over hundreds to thousands of years, with an attendant global sea level rise of around 3 m (Feldmann & Levermann, 2015). On the other hand, and perhaps more important for predicting near-future developments, rapid changes can be triggered suddenly once a longer-term process reaches a critical point. A dramatic example occurred in 2002, when 2300 km^2 of the Larsen B ice shelf collapsed in a single week (Rack & Rott, 2004) after decades of progressive ice thinning (Shepherd et al., 2003).

The Ross and Filchner-Ronne ice shelves provide a more globally significant example of a shorter-term process dependent upon a longer-term one. These ice shelves, the largest in the world, currently buttress glacier outflows from the WAIS, effectively blocking rapid ice discharge from the ice sheet into the ocean. Compared with the Amundsen Sea ice, these shelves are relatively stable; they are likely to remain in place for well over a century (e.g., Pfeffer et al., 2008; Joughin & Alley, 2011). However, after their disappearance, the WAIS could disintegrate rapidly, raising sea levels by over 3 m at rates of 1 m per century or more (Oppenheimer, 1998). In terms of timing, this process is the mirror image of the Amundsen Sea deglaciation: in the Amundsen Sea case rapid ice shelf decay is predicted to lead to a slower collapse of the ice sheet; conversely, the larger ice shelves are decaying more slowly but will allow faster ice sheet collapse once they are removed. The actual fate of the WAIS is likely to involve contributions from both processes, but their complementary mechanisms are a reminder that forecasts must consider not only individual processes but also their interactions across different timescales.

Regardless of the relative contributions from different areas or of the precise extent, it seems inevitable that the current mass loss of the WAIS (e.g., Rignot et al., 2008) will continue through the 21st century. There is still much uncertainty as to the likely extent of ice loss and the consequent contribution to global sea level change. During this century, outflow from the WAIS is likely to be at least partially offset by increased Antarctic precipitation due to atmospheric warming, which increases the moisture capacity of the air (e.g., Bengtsson et al., 2011). The Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (Church et al., 2013) estimates a likely sea level rise of −0.06 to 0.12 m (median 0.04 m) by 2100 for the RCP8.5 (highest CO_2 emission) scenario, and gives very similar estimates for their other scenarios. However, an IPCC "likely" assessment corresponds only to a 66% or greater probability, leaving a considerable chance that the Antarctic contribution to sea level change will fall outside that range. In addition, this estimate assumes that marine-based sectors of the Antarctic ice sheet will not begin to collapse before 2100; Church et al. (2013) considered the additional sea level rise from such a collapse impossible to quantify precisely, and even their suggested upper bound of several decimeters is given with only "medium confidence," corresponding to an "about 5 out of 10 chance" in the IPCC confidence level definitions. Church et al. considered it likely that such a collapse will not occur by 2100, but again only with medium confidence.

In summary, there are still major unknowns surrounding future Antarctic contributions to global sea level in the 21st century, and the IPCC's likely median estimates of 4–5 cm could be exceeded by an order of magnitude if Antarctic marine-based ice begins to collapse.

9 Conclusions and future research

Antarctica occupies a unique position: it is the planet's only polar landmass, isolated by the Southern Ocean, yet exerts a huge influence on the climate evolution of the entire planet via the growth and decay of enormous ice sheets. Because of this influence, the history of Antarctic climate not only is intriguing in its own right; it is integral to an understanding of the past and future development of the global climate system.

The modern era of Antarctic climate research began with the International Geophysical Year of 1957–8 (Florindo et al., 2008). Since then, great advances in our understanding of Antarctic climate history have been made using proxy records in ocean sediments, direct sedimentary records from the continent itself, and continually improving climate and ice sheet models. Nevertheless, many aspects of Antarctic climate and the way in which it influences and is influenced by global climate remain only partially understood—not least the details of the complex interplay among orbital forcing, the carbon cycle, and glaciation. Deep-sea drilling has provided, and continues to provide, a wealth of proxy climate data, but the recovery of direct sedimentary records from the Antarctic margins remains a challenging task. Although there are still few well-dated Antarctic cores with high recovery, they are a vital resource for corroborating deep-ocean records and model results using direct observations. As a result, every new Antarctic drilling project significantly expands our understanding of the continent's climatic evolution.

Antarctic ice and sediment coring efforts continue: the International Partnerships in Ice Core Sciences "Oldest Ice" project (Fischer et al., 2013) plans to retrieve a core that would extend the ice record from 800 kyr ago to 1.5 Ma; a third ANDRILL project is aimed at the recovery of an Eocene-to-early Miocene record from a moving ice platform at the Coulman High, McMurdo Sound (Rack et al., 2012); and a proposed IODP Leg (Escutia et al., 2013), tentatively scheduled for 2018, targets Cretaceous-to-Pliocene sedimentary records from the George V Land and Adélie Land shelf. Meanwhile, researchers continue to gain new insights from the cores retrieved by previous projects (e.g., Griener et al., 2015; Tauxe et al., 2015). Efforts at closer integration between numerical models and data from sedimentary archives (e.g., Harwood et al., 2013) are helping to produce both more-realistic models and more-reliable extrapolation from field data. An ever-growing impetus for these developments comes from the urgent need to improve projections of future Antarctic ice sheet behavior—and consequent sea level change—in response to the current anthropogenic increase in atmospheric carbon dioxide concentration. A recent priority-setting exercise for the Antarctic science community (Kennicutt et al., 2015) produced a list of vital questions about ice and ocean changes in the coming decades and centuries that can only be answered with a better understanding of the remote past: the history of Antarctic climate provides a vital key in predicting our planet's uncertain future.

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