Characterising ice sheets during the Pliocene: evidence from data and models

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Abstract

The Pliocene (~ 5.3 – 1.8 Myrs B.P.) was the last epoch of geological time in which global temperatures were generally higher than modern. It is important if we are to understand the dynamics of warm climates. This is particularly true of the interaction of climate and cryosphere, where the Pliocene may represent the first epoch in which ice sheets, at least on Antarctica, were a permanent feature. In this paper we review the available evidence for the state of ice sheets during the Pliocene as well as previous attempts to model them. We then present new models and sensitivity studies of the mid-Pliocene East Antarctic Ice Sheet (EAIS) and consider the implications for the debate on ice sheet stability during the Pliocene. These new reconstructions suggest that the mid-Pliocene EAIS was significantly smaller than modern, but the modelled average mid-Pliocene climate is not sufficient to cause the widespread deglaciation suggested by Sirius Group diatom evidence.

Keywords: *Pliocene; ice sheets; climate; models; Antarctica.*

Understanding the role of the cryosphere within the climate system is important if the mechanisms of climate are to be fully understood and future climate predictions well-constrained. Ice sheets play an important role in the modern climate, having direct and indirect influences on atmospheric temperatures, wind patterns, surface albedo, vegetation, continental water balance, ocean temperatures, sea ice formation and ocean circulation patterns (Clark *et al.* 1999). However, discussion continues over their significance in deep time palaeoclimates. Whether ice sheets are one of the primary drivers of Cenozoic climate change or a largely passive responder to other changes in the climate system remains an open question.

After the initial transition into a world with significant glaciation at ~34 Ma, the waxing and waning of both Northern Hemisphere and Antarctic ice sheets has been used to explain the variation of a large variety of Cenozoic palaeoclimate proxies, particularly fluctuations in marine $\delta^{18}O$ oxygen isotope ratios (Kennett 1977; Matthews 1986; Zachos 2001). However, there are no direct proxies for ice volume, since, for example, $\delta^{18}O$ values also depend on bottom water temperatures and ocean salinity (Rostek *et al.* 1993). Even when assumptions are made which allow a global ice volume to be inferred, there is typically no method to identify the volume or location of individual ice sheets. This ambiguity is further complicated by the non-linearity of climate, ice sheet dynamics and the interactions between the two. In a system as complex as the ice sheet – climate system, it is not always immediately obvious what the observed values and fluctuations of climate dependent variables represent. In order to reconstruct the ice sheets from deep time periods of the Earth's history, and thus investigate glaciations in warm climates, it is important that direct evidence of ice sheet extent is collected in order to constrain models. These models can in turn interpret the consequences of this evidence and reconstruct the ice sheet as a whole.

The Pliocene epoch is important, as it is the last time when the Earth was warmer for a period longer than any Quaternary interglacial, providing a unique window into the workings of the ice sheet - climate system. Deep time cryospheric data is limited within the geological record, however even limited data coverage can constrain models and provide important boundary conditions. In this paper we review existing ice sheet data and models, showing that much can already be inferred regarding Pliocene ice sheets, whilst debate continues over some of these interpretations. We then present new model reconstructions of the East Antarctic Ice Sheet (EAIS) and consider the implications of the results.

Importance of the Pliocene

If we are to understand the future of our climate, we need to understand the past Earth system and its potential modes of operation. Greater understanding of how the climate will respond to increased greenhouse gas concentration and how transitions, from one climate state to another, occur is required. These required areas of understanding necessitate the study of palaeoclimates. Of particular interest to those looking at future climate change are periods with increased temperatures, greater concentrations of greenhouse gases in the atmosphere and periods in which large climate transitions occurred.

Cenozoic records show that climate has cooled from the hot, greenhouse worlds of the Palaeocene and Eocene, through a number of gradual and stepwise transitions to the bipolar glaciations of the Pleistocene (Miller *et al.* 1987; Lear *et al.* 2000; Zachos *et al.* 2001). The Pliocene is crucial to our understanding of the response of ice sheets to these changes in climate. It is believed that large scale, permanent Antarctic glaciation was established in the middle Miocene (Flower & Kennett 1994; Zachos *et al.* 2001), whilst the first indication of limited Greenland glaciation occurs in the late Miocene (Larsen *et al.* 1994). This means that the late Miocene and Pliocene are a unique Cenozoic testing ground for the interaction of large scale Antarctic and limited Northern Hemisphere ice sheets with a warmer than modern climate. If the nature and extent of the Pliocene cryosphere can be determined then this could help us to understand the way the ice sheets, and therefore climate, may respond to future global warming.

The Pliocene epoch also has great significance in a wider understanding of climate change, not least because it is 'geologically recent', providing large quantities of palaeoenvironmental data. The reliability of these data, compared with previous warm periods, is increased by improved geographical distribution, biota-environment correlations and stratigraphy (Dowsett *et al.* 1992). The mid-Pliocene is believed to be the last time in Earth history when global temperatures where sustained at levels higher than today. A number of geological (Dowsett *et al.* 1999) and modelling studies (Sloan *et al.* 1996; Haywood *et al.* 2000; Haywood & Valdes 2004) indicate that mean temperatures during this interval were between 1.4 and 3.6°C higher than present. After this warmth the climate cooled and underwent a transition that saw the onset of widespread Northern Hemisphere glaciation, beginning approximately 2.7 Myr B.P. in the late Pliocene. Thus the Pliocene is the last interval at which temperatures were sustained at levels equating to those predicted for the coming century (Fig. 1). Unfortunately, the concentration of CO_2 in the Pliocene atmosphere is only partially constrained. Some proxies indicate that levels were similar to, or slightly higher than, those measured today (Van der Burgh *et al.* 1993; Raymo *et al.*

1996), whilst others suggest values closer to the Last Glacial Maximum (Pearson & Palmer 2000). These indirect proxies predict similar values for the Pleistocene and Eocene, but have very different results for the Pliocene. This suggests that the calibration of one or more of these may not reflect the true sensitivity of the proxy to atmospheric carbon.

Although there is a consensus that the global climate of the Pliocene was significantly warmer than modern, the mechanisms that caused the enhanced temperatures and the pattern of warming have yet to be definitively determined. The polar amplification of temperatures, combined with little, if any, tropical warming in mid-Pliocene sea surface temperature reconstructions suggest that enhanced oceanic poleward heat transport could be the major contributing factor (Dowsett *et al.* 1992; Kwiek & Ravelo, 1999). However, coupled ocean-atmosphere General Circulation Models (GCMs) fail to reproduce the proposed increase in meridional heat transport and suggest significant tropical warming (Haywood & Valdes 2004), more indicative of CO_2 induced heating. However, it remains unclear whether the discrepancy between the data-based reconstructions and the models arises from insufficient data coverage, lack of proxy sensitivity, errors in the GCM boundary conditions or inaccuracies in the physics of ocean circulation changes. Recent data using alkenone palaeothermometry techniques supports the model results, showing enhanced temperatures in the equatorial Pacific and Atlantic (Haywood *et al.* 2005; Williams *et al.* 2005).

A modelling study aimed at establishing the relative contributions of the various mechanisms for producing enhanced global temperatures found that the most significant forcing was provided by the smaller prescribed ice sheets (Haywood & Valdes 2004). The PRISM (Pliocene Research, Interpretation and Synoptic Mapping) mid-Pliocene environmental reconstructions incorporate a 50% reduction in Greenland ice volume and a 33% reduction in Antarctic ice volume (Dowsett *et al.* 1999). These ice sheet volume reconstructions were based on sea level and marine oxygen isotope data that have large error bars associated with them (Lietz & Schmincke 1975; Dowsett & Cronin 1990; Krantz 1991; Wardlaw & Quinn 1991). These error bars encompass vastly different global ice sheet scenarios, from significant deglaciation of East Antarctica, to all the sea level increase originating from Greenland and West Antarctica.

Pliocene Northern Hemisphere Glaciation

The Plio–Pleistocene climate transition, ~ 2.7 Myrs B.P., principally records the development of large ice sheets over Fennoscandinavia and North America, which probably had not previously existed during the Cenozoic. However, the Greenland Ice Sheet (GrIS) is more resilient to warmer temperatures and a large portion probably survived the, warmer than modern, last interglacial (Otto-Bleisner *et al.* 2006; Overpeck *et al.* 2006). Small amounts of ice rafted debris (IRD) have been found in Greenland continental shelf records for the last 7 million years (Larsen *et al.* 1994), but this is not necessarily indicative of large ice sheets in Greenland. Mountain glaciers can produce IRD when they reach down to the coast, requiring only small total ice volumes. Indeed, the first significant volume of IRD on the Greenland continental margin is observed at approximately 3 Ma, which may represent the first major increase in Greenland ice volume (Jansen *et al.* 2000).

Although the Neogene geological record on Greenland is sparse there is some evidence that the GrIS was significantly smaller than at present during the Pliocene epoch. The shallow marine deposits from Ille de France in north-east Greenland indicate summer temperatures more than 6°C warmer than today during the mid-Pliocene warm period (Bennike *et al.* 2002). Furthermore, evidence for forestation in the northernmost reaches of Greenland, as late as 2.4 Myrs B.P. is found at the Kap Kopenhavn formation (Funder *et al.* 2001) and willow and pine fragments in the basal ice of the NGRIP (North Greenland Ice core Project) ice core suggests forestation in central Greenland at the point of glacial inception (Dahl-Jenson 2006).

Pliocene deposits across the Canadian Arctic islands, northern Alaska and northern Russia show that the Arctic was generally warmer in the Pliocene and that the tree line had migrated well into the Arctic Circle (Fig. 2). Particularly telling is the evidence for evergreen forests on Meighen (Matthews 1987) and Ellesmere islands

(Fyles 1989), today a polar desert. The terrestrial biota found there and around the Arctic suggests that large-scale glaciation in the Pliocene Northern Hemisphere could only have existed on Greenland, with possible localised glaciation on Ellesmere Island (de Vernal & Mudie 1989) and in Alaska (Lagoe & Zellers 1996). The available data from Greenland are indicative of a significant reduction in the extent of the ice sheet during the Pliocene.

Pliocene Antarctic Peninsula Ice Sheet

The Antarctic Peninsula Ice Sheet (APIS) is the smallest and most northerly component of Antarctic ice volume. It is also the area of the Antarctic that has shown most response to recent climate changes (Vaughan *et al.* 2003; Cook *et al.* 2005). A number of ice shelves have collapsed probably due to the formation of surface melt features, as warmer summer temperatures propagate further south (Vaughan & Doake 1996; Scambos *et al.* 2000). The removal of this buffer has been shown to cause the retreat of grounding lines and acceleration of the glaciers that once fed the ice shelf (De Angelis & Skvarca 2003; Rignot *et al.* 2004). Owing to this high degree of climate sensitivity many researchers have assumed that the peninsula would generally be ice free in the warmer pre-Quaternary Cenozoic palaeoclimates (Abreu & Anderson 1998; Dowsett *et al.* 1999).

However, recent evidence has shown that significant ice must have existed on the Antarctic Peninsula for large portions of the post-Eocene Cenozoic (Dingle & Lavelle 1998; Ivany *et al.* 2006). Marine geological and geophysical studies of the Antarctic Peninsula Pacific margin, at Ocean Drilling Program (ODP) sites 1095, 1096, 1097 and 1101 (Fig. 3), seem to show that ice must have been present during the Neogene. Seismic reflectors, which have been interpreted as glacial unconformities, give evidence for a large APIS as long ago as the middle Miocene. Correlation with biostratigraphic dating of ODP Leg 178 cores, allowed a date of between 5.12 and 7.94 Ma to be assigned to a series of unconformities, suggesting that the APIS advanced out onto the continental shelf at least 12 times during the late Miocene (Bart *et al.* 2005). The number of unconformities seems to increase in the Pliocene, with 30 recorded by Bart & Anderson (2000), but this may represent preservation rather than intrinsic rate of occurrence. The Antarctic Peninsula frequency was compared to similar sites in the eastern and western Ross Sea, which was speculated to show that the APIS was more dynamic during the late Neogene than either the West Antarctic Ice Sheet (WAIS) or EAIS, although preservation issues and accurate dating of the seismic sequences remain unresolved. The advance of the APIS over the continental shelf is substantiated by the occurrence of Pliocene sediments in large drift deposits located on the adjacent continental rise. They indicate repeated advance and retreat of grounded ice masses across the shelf and are associated with IRD of Antarctic Peninsula provenance (Cowan 2002; Hillenbrand & Ehrmann 2002, 2005; Pudsey 2002). Physical, geochemical records and X-ray images, derived from ODP Leg 178 Site 1095, show variations attributed to glacial – interglacial cycles during the early Pliocene (Hepp *et al.* 2006).

The limited terrestrial record seems to support the picture of a fluctuating Pliocene APIS. Deposits, interbedded with igneous rocks, at Cockburn Island (Dingle *et al.* 1997) and James Ross Island (Kristjansson *et al.* 2005; Hambrey & Smellie 2006; Smellie *et al.* in press), are believed to represent at least periodic interglacial conditions, with faunal assemblages indicative of warmer than modern marine conditions (Zinsmeister & Webb 1982) and little evidence of IRD (Jonkers *et al.* 2002). Other Pliocene volcanism on the Antarctic Peninsula gives further evidence of a complex cryospheric history with both englacial and subaerial eruptions at Seal Nunataks (Smellie & Hole 1997) and Hornpipe Heights (Smellie 1999) respectively.

Pliocene West Antarctic Ice Sheet

The WAIS is currently the only marine ice sheet (one that is largely grounded on bedrock below sea level). It is particularly difficult to define its history prior to the Last Glacial Maximum, as erosion of the surrounding continental shelf has removed all surface evidence of previous glaciations on the seabed. Few rock outcrops occur in the interior of the ice sheet, making any evidence of pre-Quaternary glaciations very difficult to obtain. However, a limited record of the state of the WAIS during the Pliocene does exist (Fig. 3).

Mount Sidley in the Executive Committee Range, Marie Byrd Land, is the highest volcano in Antarctica. The outcrops are dominated by subaerially erupted volcanic rocks, hence the most recent eruptions must have occurred when there was little ice cover in the region. These rocks have been dated to between 5.7 and 4.2 Ma, suggesting the early Pliocene WAIS was reduced in size compared to today (Panter *et al.* 1994). Further Pliocene dates are found on another West Antarctic volcano, Mount Murphy, which is located on the coast, beside the Crosson Ice Shelf. Mount Murphy has an unusual association of volcanic rocks, glacially derived sediments and recycled microfossils and a varied ice sheet history (LeMasurier *et al.* 1994; Smellie 2000). A similar scenario has been proposed for these sediments as for the Sirius Group deposits in the Transantarctic Mountains (see later), meaning the diatoms found there could represent periods of marine deposition in the interior of West Antarctica. These sediments are of particular note in this discussion due to the presence of diatoms restricted to the Pliocene in the Southern Ocean, suggesting that the Byrd Subglacial Basin, to the south and upstream of Mount Murphy, may have been ice free during some of the Pliocene (LeMasurier *et al.* 1994).

Attempts to extract microfossils from underneath the currently grounded WAIS have been undertaken in the Siple Coast region. During a campaign to drill through the Crary Ice Rise, at the mouth of Whillans Ice Stream (formerly Ice Stream B), sediment was removed from under the ice dome. The diatoms recovered were from the late Miocene, with no diagnostically post-Miocene species (Scherer *et al.* 1988). This does not necessarily mean that marine deposition stopped at the end of the Miocene, as the ice rise is believed to have only been grounded for around 1100 years (Bindschadler *et al.* 1989), but the sediments above the Miocene layers may have been subsequently eroded by the WAIS. Microfossils were also found in the sediments underneath the upstream section of the Whillans Ice Stream. Diatoms from the middle Miocene to the present were found, with the temporal ranges of the species potentially showing a continuous deposition record. However, none of the species are diagnostic of a Pliocene age, whereas some of the diatoms are limited to the Miocene and Pleistocene, suggesting both these periods saw deglaciation of central West

Antarctica (Scherer 1991). It is hard to draw specific conclusions about the Pliocene, but if the WAIS grounding line can retreat beyond the upstream sections of the Ross Sea sector during both the late Miocene and the Pleistocene, then this may also be expected in the Pliocene.

Marine evidence for the state of the WAIS is available from the Ross Sea and Weddell Sea regions. Neogene glacial unconformities have been observed in seismic stratigraphic studies of the eastern and western continental shelf of the Ross Sea. These are thought to represent periods of sufficient ice sheet expansion to cause the grounding line to advance and predicate subglacial erosion. Both regions were found to have 8 unconformities during the Neogene, which suggests at least 8 grounding events, representing expansions of the EAIS and WAIS respectively (Bart & Anderson 2000). This would seem to show that, at least during the Neogene, both the EAIS and the WAIS experienced fluctuations and were prone to large scale changes in volume.

Beyond the Weddell Sea shelf margin, continuous deposition occurred throughout the late Miocene, Pliocene and Pleistocene (Barker & Kennett 1988). In fact the Pliocene saw faster sedimentation than the Pleistocene, under conditions that, looking at other evidence from marine sediments, should have seen a highly variable sediment supply, as the ice sheet advanced and retreated across the Weddell Sea continental margin. It could be argued that during this period the Weddell Sea locality must have been a stable, heavily glaciated environment, in order to continuously supply the ocean basin with sediment. However, a further scenario has been proposed, where sediment is supplied by local glaciers transporting material through deep canyons, bypassing the continental shelf and negating the need for large-scale glaciation in the region (Barker 1992).

Pliocene East Antarctic Ice Sheet

The behaviour of the EAIS is the most contentious aspect of the Pliocene ice sheet stability debate. The two scenarios that have been proposed are a dynamic ice sheet, experiencing large scale deglaciations throughout the late Miocene and Pliocene and a continent wide cold based EAIS, since the inception of a permanent Antarctic ice sheet in the middle Miocene (Fig. 4). The two hypotheses and their supporters, daubed ‘dynamicists’ and ‘stabilists’, have been at loggerheads for more than 20 years and the questions they raise still lack completely satisfactory answers.

The controversy began with the discovery of Sirius Group marine diatoms high in the Transantarctic Mountains. The mechanism originally proposed for the emplacement of these diatoms is glacial transport in times of an expanded EAIS from their marine deposition sites in the modern day Wilkes-Pensacola Sub-glacial Basins. The location of the tills in which diatoms are found would require the almost complete deglaciation of the basins, which could correspond to a rapid ice volume reduction of up to 60%. The chronology of these events has been approximated by dating the diatoms found in the Sirius Group sediments (Fig. 3) and forms a discrete set of marine incursions in the Miocene and Pliocene (Fig. 4).

Evidence of grounding line retreat from the Lambert Glacier region supports these interpretations. Miocene and Pliocene marine deposits, subsequently uplifted above the modern ice surface in the Prince Charles Mountains, hundreds of kilometres inland from the modern grounding line, seem to correlate well with the deposition periods of the Sirius Group diatoms. The deposition of the Pliocene member of the group, the Bardin Bluffs sediments, has been biostratigraphically dated as occurring between 3.4 and 2.6 Ma (Whitehead *et al.* 2004). The presence of three planktonic taxa in reasonable abundance points to the existence of ice free marine surface waters when the Bardin Bluffs sediments were deposited (McKelvey *et al.* 2001). Micropalaeontological studies of ODP sites 1165 and 1166, in Prydz Bay, show that the sea ice in the region was reduced by up to 78% during the Pliocene (Whitehead *et al.* 2005). There is also evidence of significantly increased early Pliocene temperatures (summer sea surface temperatures >3°C) and open water conditions at the nearby Sørsdal Formation (Whitehead *et al.* 2001).

The Beardmore Glacier region also provides some possible evidence of a very different East Antarctica during the Pliocene. The Oliver Bluffs deposits, suggested as mid-Pliocene in age by the diatom assemblage, contain exceptionally preserved examples of *Nothofagus* (Southern Beech) leaf mats, twigs and wood (Francis & Hill 1996). Ecological interpretation of these and other Oliver Bluffs finds, which include palaeosols, palynomorphs, mosses, liverworts, conifers, angiosperms and two species of weevil (Askin & Markgraf 1987; Retallack *et al.* 2001; Ashworth & Kuschel 2003; Ashworth & Cantrill 2004), suggest a warmer than modern, polar environment, with mean annual temperatures of approximately -8°C. The use of the diatoms to date these sediments is controversial. However, provenance studies of the Sirius Group deposits suggests the Oliver Bluffs may be one of the youngest members, as they seem to represent erosion of the local glacial troughs (Passchier 2004). This finding, along with marine evidence of *Nothofagus* in Pliocene sediments in the outer Ross Sea (Fleming & Barron 1996), supports the existence of localized dwarf *Nothofagus* forests on Antarctica, until relatively recently.

The assertion of a dynamic EAIS is disputed by the ‘stabilists’, who maintain that there has been little change in East Antarctica since the middle Miocene, when large scale glaciation was established. This hypothesis was established with the compilation of Cenozoic $\delta^{18}O$ records, which showed a clear shift towards heavier values (colder climate and/or greater global ice volume) in the middle of the Miocene (Kennett 1977; Zachos *et al.* 2001). The terrestrial evidence for this view comes mainly from the remarkable landscape stability of the Dry Valleys region of the Transantarctic Mountains. Geomorphological studies (Sugden *et al.* 1995) and ashfall dating techniques (Marchant *et al.* 1996) have been used to infer that the region has been essentially unchanged, climatically (Sugden *et al.* 1999; Summerfield *et al.* 1999), tectonically (Fitzgerald & Stump 1997; Ackert & Kurz 2004) and glaciologically (Stroeven & Prentice 1997; Goff *et al.* 2002), since the Miocene. Some researchers have suggested that the Dry Valleys region is glaciologically unique, and could have seen little change during periods of major deglaciation elsewhere in East Antarctica (Kerr & Huybrechts 1999). However, the level of landscape stability reported is difficult to reconcile with a scenario of EAIS dynamism and significant Antarctic warming. This gives credence to alternative suggestions, such as atmospheric transport (Burckle & Potter 1996; Kellogg & Kellogg 1996; Stroeven *et al.* 1996) and meteorite impact fallout (Gersonde *et al.* 1997), for the emplacement of the Sirius Group diatoms.

The Pliocene marine record around East Antarctica is limited and, in many cases, poorly dated. However, there are a few coring expeditions that have obtained Pliocene material. At opposite sides of the continent, at Ferrar Fjord (CIROS-2 drill site), Taylor palaeofjord (Dry Valleys Drilling Program (DVDP) holes 10 and 11), Prydz Bay (ODP Site 1165) and Gunnerus Ridge (Polarstern cores PS1811-8 and PS1812-6), sediments appear to alternate between glacial and interglacial type deposition (Barrett *et al.* 1992; Ishman & Reick 1992; Hillenbrand & Ehrmann 2003; Warnke *et al.* 2004). Combined with evidence of erosional unconformities in the Ross Sea sediments (Bart & Anderson 2000), this suggests that the EAIS fluctuated, at least at the margins, during the Pliocene.

Clearly the history of the EAIS is a complex one and it seems increasingly likely that the ice sheet does not generally behave uniformly. The individual drainage basins (Vaughan *et al.* 1999) and tectonic provinces (Fitzgerald *et al.* 1986; Fitzgerald 1994; Hindmarsh *et al.* 1998) could have responded very differently to climatic changes. Establishing exactly what this response has been to the various known climate events is severely hampered by the lack of available terrestrial evidence. It is in response to this scarcity of evidence and the differing interpretations of the existing geological record that modelling of the EAIS during warm periods of the past, such as the Pliocene, must test the different paradigms of ice sheet behaviour.

Modelling Ice Sheets in Warm Climates

There is a reasonable amount of information on the state of the ice sheets during the Pliocene, but this has not been sufficient to finally settle many of the debates regarding the cryosphere. In such a situation modelling should allow the gaps in the data to be filled in. However, accurate ice sheet modelling requires high-resolution boundary conditions for any particular time period. Ice sheet models (ISMs) are generally driven by temperature, accumulation and melting, as well as being heavily dependent on the bedrock on which the model is run. Clearly, extracting such parameters from the geological record, on a grid of high enough resolution to be useful within these models (e.g. at 50 km intervals), for time intervals before satellite observations are available is extremely difficult.

There are a number of potential ways to overcome this problem. One of the ways utilised in previous studies is to use modern observations or a modern parameterisation as a basis from which to increase temperatures and examine the effect these increases have on the ice sheets. Huybrechts (1993) used a 3-D ISM of Antarctica to test the effect of a constant increase in surface temperature, up to 20°C above modern. Temperature increases of less than 5°C were found to increase East Antarctic ice volumes, owing to the dominance of a parameterised increase in precipitation over melting, modelled by the positive degree-day (pdd) method. Warming of greater than 15°C was required to deglaciate the Wilkes-Pensacola Sub-glacial Basins, as hypothesised by Harwood & Webb (1986; Webb & Harwood 1991; Harwood 1983). It was considered that 15°C of temperature increase was probably more than would be expected in late Tertiary warm climates (Huybrechts 1993).

More sophisticated modelling has allowed the potential errors caused by insufficient knowledge of the boundary conditions to be examined. Huybrechts & Kerr (1999) undertook similar experiments to those of earlier work (Huybrechts 1993), but examined the effect that changes in the elevation of individual tectonic blocks within the Transantarctic Mountains had on ice surface elevations. The tectonic history of the Transantarctic Mountains is poorly understood and contributes significantly to the uncertainty in reconstructing Antarctic palaeogeography. Evidence exists that some areas have been lower in the recent past (Behrendt & Cooper 1991; McKelvey *et al.* 1991; Webb *et al.* 1996; Hambrey *et al.* 2003). However, the region has experienced a complex history, with the Transantarctic Mountains split into a number of tectonic blocks (Fitzgerald *et al.* 1986; van der Wateren *et al.* 1999) that may have undertaken very different movements (Fitzgerald 1994; Hindmarsh *et al.* 1998). The study of Huybrechts & Kerr (1999) reveals that tectonic changes can have a large effect locally, but do not seem to cause significant long range changes in ice sheet elevation. This result could hamper attempts to compare ISM reconstructions with available data, which often tends to relate to single localities. This is particularly true in the Dry Valleys region, which is key in the arguments between the ‘dynamicists’ and ‘stabilists’. In this region a modelled temperature increase of 10°C caused the EAIS margin to retreat significantly, but a local ice cap remained on the Dry Valley mountaintops. This suggests that the exceptional stability observed could reflect the local glaciological and climatic conditions, with inferences and conclusions drawn from the region not reflecting the wider EAIS behaviour.

Modelling mid-Pliocene Ice Sheets

Although there is much to be learnt from previous sensitivity studies, their applicability to specific ice sheets within the geological record is uncertain. This is because of the assumptions implicit within these models. Ice sheets experience altitude-temperature, ice-albedo, sea ice and ocean circulation feedbacks as well as internal non-linearities, and the results are therefore highly dependent on the boundary and initial conditions, particularly the ice surface height that the models are initiated with. None of the studies described above have explored the effect that initial conditions would have on the conclusions. A further assumption implemented in the application of changes in temperature is that Antarctic temperatures increase smoothly, with little regional variation. However, the climate is highly non-linear and changes to the circulation and temperature patterns could be significant in warmer climates of the past (Rial *et al.* 2004). It is therefore important that more sophisticated climate modelling of Antarctica in warm periods of the geological past is applied to the boundary condition of ISMs.

The two existing models utilised in this study are the British Antarctic Survey ISM (BASISM) (Hindmarsh 1999, 2001) and the United Kingdom Met Office (UKMO) GCM, specifically the HadAM3 atmospheric GCM. The atmospheric component of the UKMO GCM consists of 19 vertical layers, with a horizontal resolution of 2.5° in latitude and 3.75° in longitude. A more complete technical summary of the UKMO GCM is described in Pope et al. (2000) in which the impacts of improvements to the model, compared to the previous Hadley Centre model, are discussed. BASISM is a thermomechanically coupled, three-dimensional ice sheet model, similar to those of Huybrechts (1990) and Ritz et al. (1997). The model uses the shallow ice approximation to close the mass, momentum and thermal balance equations that govern ice flow, and describe the grounded ice component. This numerical scheme has yet to be extended to include the more complex flow regimes of ice streams and ice shelves, as the understanding of the physics in these regimes is incomplete.

The ISM was run on a 50km × 50km polar stereographic grid, in a domain covering the modern grounded EAIS. The GCM is run on a 2.5° lat × 3.75° long grid and hence output climate fields require downscaling onto the ISM grid. The relevant fields from the GCM model were downscaled using a bilateral interpolation technique and simple lapse rate and albedo corrections. The pdd method was then employed to convert these climate fields into an accumulation / melt rate (Reeh 1991; Braithwaite 1995). This technique assumes that the melting of the ice sheet surface can be fully described by three physical constants and the temperature record, which, although many other factors could contribute, has been shown to have some physical justification (Ohmura 2001).

The mid-Pliocene EAIS can thus be modelled by taking the output of extant mid-Pliocene GCMs and driving BASISM with the climatological output fields. Rather than previous modelling, that essentially tested the sensitivity of the modern ice sheet to uniform temperature changes, a realistic mid-Pliocene climate can now be reproduced and the ice sheet, during this particular period of time, reconstructed from the modelled climatology (Fig. 5) and BEDMAP topographic reconstruction (Fig. 6). Using an ISM in conjunction with the GCM allows us to attribute the ice sheet reconstruction to this particular period of the geological record and hence test the results against the geological record. The employed modelling strategy also allows the effects of ice sheet hysteresis to be explored, rather than assuming the ice sheet deglaciated from the modern configuration. The ‘standard’ mid-Pliocene climate reconstruction is a HadAM3 experiment (Haywood *et al.* 2000), whose boundary conditions are based on the environmental reconstructions of the PRISM group (Dowsett *et al.* 1999). These are designed to represent the values averaged over the mid-Pliocene time slice (3.29-2.97 Ma).

The models depicted in Figure 7 are the end members of a suite of EAIS reconstructions, under the climate forcing of the 'standard' mid-Pliocene GCM experiment. The difference between the reconstructions is due to the feedback induced hysteresis in the ice sheet – climate system. The coupling between the GCM and ISM is offline, meaning the GCM experiment is not rerun with the BASISM predicted ice sheets. This eliminates a number of important feedbacks; however, altitude-temperature and ice-albedo feedbacks are modelled and account for the differences between ISM results.

The models show that there were significant differences between the modern and mid-Pliocene ice sheets on Antarctica. Modern Antarctic temperatures rarely climb above 0°C, especially in East Antarctica (Comiso 2000), therefore very little ice sheet runoff is observed or predicted (Jacobs *et al.* 1992). This is not the case in the mid-Pliocene climate simulations, where average January temperatures are significantly above the melting point of ice across some parts of East Antarctica (Fig. 5b). Hence, surface melting can become a significant factor in the equations of ice sheet mass balance and drive deglaciation of the coastal regions of Antarctica (Fig. 7c). This type of ice sheet – climate regime is today restricted to Greenland, where surface melting constitutes at least 50% of all ice loss (Hanna *et al.* 2005).

In all cases the extent of the ice sheet is reduced in the Wilkes and Aurora subglacial basins, however the average mid-Pliocene climate, modelled with the GCMs, seems insufficient to cause the large-scale deglaciation suggested by the diatom evidence. Although, in the lower ice volume reconstructions, the northern reaches of the Wilkes Subglacial Basin are deglaciated, this does not extend southwards to some of the drainage basins from which the proposed transport of marine diatoms into the central Transantarctic Mountains occurred (Harwood & Webb 1986).

Clearly there are uncertainties in the boundary conditions that are utilised in these models. In order to test the effect of two of the most important boundary conditions, surface temperature and the bedrock topography of the subglacial basins, on the conclusions drawn from the equilibrium ice sheet reconstructions sensitivity experiments were undertaken. Antarctic temperatures are notoriously difficult to reproduce, either within a GCM modelling context (Connolley & Cattle 1994) or from observational datasets (Comiso 2000). The potential errors in modelling mid-Pliocene Antarctic temperatures are only increased by the lack of palaeoenvironmental data and could introduce significant uncertainty to the Pliocene ice sheet model reconstructions. Thus, in order to test the sensitivity to the modelled temperature, the 'standard' mid-Pliocene GCM temperatures were incremented in discrete steps up to 10°C. The large magnitude of this temperature anomaly should also encompass the potentially large, orbitally driven, temperature changes that occur within the PRISM defined mid-Pliocene time slab (Haywood *et al.* 2002).

Some areas of the Antarctic have been explored extensively in recent years and many now have good coverage of bedrock topography data. However large parts of the continent, particularly in East Antarctica, are unexplored and have little or no data on the subglacial environment (Lythe *et al.* 2001, p. 11338, Figure 2b). Although a limited geological history of Antarctica can be reconstructed from the outcrops at the ice sheet margin, the tectonic history of the Antarctic interior since the Pliocene is largely unknown. There are three main mountain ranges in East Antarctica. The geological history of the subglacial Gamburtsev Mountains is essentially unknown (Dalziel 1992), although it has been suggested they may have formed during the Carboniferous (Veevers 1994). The mountains of Dronning Maud Land are a series of ancient mountain belts, believed to have formed during the middle to late Mesozoic (Näslund 2004). Indeed, evidence from cosmogenic dating of glacial tills in the Sør Rondane Mountains shows Pliocene ages for tills >300m above the present ice surface (Moriwaki *et al.* 1992), suggesting this region had sufficient topography to benefit from the increased precipitation in the warmer Pliocene, without significant melting. The Transantarctic Mountains are the most well studied of East Antarctica's mountains, but provide an altogether more complex picture. Their Tertiary and Quaternary history is not well established and differing lines of evidence produce very different conclusions (Behrendt & Cooper 1991; Brook *et al.* 1995; Sugden *et al.* 1999; van der Wateren *et al.* 1999). This would seem to add a large uncertainty in the boundary conditions for our ice sheet models. However, previous modelling studies seem to indicate that changes to the Transantarctic Mountain altitudes have only local consequences, with little impact across the ice sheet as a whole (Kerr & Huybrechts

1999). Altogether this evidence suggests that the East Antarctic mountain topography does not add significant uncertainty to Pliocene ice sheet modelling, however the same may not be true of the subglacial basin topography.

The altitude of the Wilkes and Pensacola subglacial basins could have a number of potential impacts on the plausibility of the Pliocene deglaciation scenarios. The greater ice depth provides an increased potential altitude – temperature feedback. It also reduces the required rapidity of the proposed deglaciation, as a deeper basin would take longer to isostatically rebound to an altitude above sea level. If it is assumed that the dynamicist hypothesis is correct then there should be a significant depth of sediment in the basins, that would not have existed during the deglaciation phase of the mid-Pliocene. Hence in our sensitivity studies the altitude of the subglacial basins (Fig. 6b) was decreased by up to 1000m, simulating the removal of possible sediment packages.

Even in the most extreme scenarios modelled here, the southernmost areas of the sub-glacial basins remain glaciated (Fig. 8), meaning that marine deposition could not have occurred here. It appears from this study that the average mid-Pliocene climate is not sufficient to explain possible Pliocene ice sheet collapse, enabling marine deposition in the southernmost Wilkes-Pensacola Sub-glacial Basin, even when tested errors in Antarctic temperature reconstructions and basin topography are taken into account. However, there remain a number of possible explanations of Pliocene marine deposition in the southern Wilkes Basin.

By analogy with the rapid deposition of sediments at the mouths of fast flowing, polythermal glaciers in East Greenland (Dowdeswell *et al.* 1994), it was estimated that the mid-Pliocene sediments at Bardin Bluffs, could represent as little as hundreds to thousands of years of marine deposition (Hambrey & McKelvey 2000). If this proposed scenario occurred in the Lambert Glacier region, then it is plausible that a similarly short collapse could have occurred in the Wilkes Basin during the warmest period of the mid-Pliocene. The climatic conditions for such a collapse would not be simulated in the existing reconstructions of the mid-Pliocene climate, which are averaged over a 300,000 year interval (Dowsett *et al.* 1999).

One of the regions that recent measurements suggest is changing most rapidly is the Pine Island region, which has the greatest observed ice thinning rate in Antarctica (Thomas *et al.* 2004; Davis *et al.* 2005). The oceans have been implicated in this rapid thinning, possibly a result of warm water flowing onto the continental shelf towards the ice stream grounding line (Jenkins *et al.* 2004; Payne *et al.* 2004; Shepherd *et al.* 2004; Bindschadler 2006). Such an ice sheet - ocean interaction could potentially lead to rapid and extensive ice sheet retreat. However, the oceanographic perturbation would probably be beyond the resolution of the current generation of GCMs and such dynamical interaction is not possible with the current ISMs.

Another factor that may have led to an overestimation of the mid-Pliocene EAIS stability is the lack of a self-consistent representation of ice streams within existing ice sheet models. Ice streams are the most dynamic of Antarctic Ice Sheet features (Rose 1979; Jacobel *et al.* 1996; Shepherd *et al.* 2002) and are responsible for over 90% of modern ice and sediment discharge (Bamber *et al.* 2000). They may also play an important role by propagating changes from the margins deep into the ice sheet's interior (Bamber *et al.* 2000; Payne *et al.* 2004). If a palaeo-ice stream existed in the Wilkes Basin, this could have a major implication on the ease with which ice could be removed from this region.

Our study only couples the GCM and the ISM in an offline scheme. This misses some of the major feedbacks in the ice sheet – climate system and as such may introduce uncertainties in the results. These potentially important feedbacks include sea ice, precipitation, atmospheric circulation, melting, iceberg calving rate, ocean temperature and circulation feedbacks. Currently there are no ice sheet models that effectively reproduce the dynamics of the sea-ice boundary. A full appreciation of the physics of grounding-line migration is only now being gained (Schoof in-press). Whereas, records of ice sheet calving and oceanic boundary melting rates are spatially limited and only extend back a few decades (Losev 1963; Jacobs *et al.* 1992; Rignot & Thomas 2002), meaning that the dependence of rates on climatic and topographic parameters is poorly understood. In this modelling study the problems presented by grounding line migration are overcome by fixing the model grounding line at the observed, modern grounding line (Lythe *et al.* 2001). Sensitivity experiments (not presented here) on calving rates (in which only a water depth dependence was included), showed that even for large rates the results presented here are not conceptually different. Although calving can cause significant drawdown of the ice sheet, the reduction in extent that this can produce is limited by the topography and isostatic response rather than the magnitude of the calving rate.

Clearly there is much work to do before a complete understanding of Pliocene ice sheets can be achieved. Terrestrial data from Antarctic exposures will always be sparse and important new revelations from this source, although possible, are probably unlikely. There is some scope for important discoveries, such as the nature of the Wilkes Basin bed, from geophysical studies of the Antarctic continent. Continental shelf marine records also have much to offer, with a far greater possibility of new Pliocene sediments coming to light. Drilling around Antarctica offers the possibility of being able to constrain the advances of the EAIS and the WAIS onto the continental shelf. However, this record will never give us the complete picture, as sediments will have been eroded away by subsequent ice advances and evidence of the extent of retreats will, naturally, be limited.

Modelling remains an important tool if we are to understand Antarctic palaeoclimates. There are many improvements that remain to be implemented if we are to be able to accurately reconstruct past ice sheets. However, advancements both in our understanding of the physics of the ice sheet - climate system and in modelling techniques continue to be made. In the near future more sophisticated ISMs will be fully integrated into climate models and careful reconstruction of the boundary conditions will allow more accurate constraint of palaeo-ice sheets.

Conclusion

The Pliocene is an important epoch for palaeoglaciology. It marks the transition from warmer climates with smaller, possibly more dynamic (Rebesco *et al.* 2006) ice sheets into the bipolar, cold, icehouse world of the Pleistocene. Evidence and modelling of the Pliocene ice sheets suggests that they existed in a generally smaller form compared to present, but fluctuated significantly. The major question that remains, and particularly so for the EAIS, is the magnitude and limits of these fluctuations. Modelling of the climate and ice sheets can help constrain these values. However, these techniques are only now being turned to the Pliocene and there remains room for further development.

Sensitivity modelling of the EAIS suggests that the climate, averaged over the ~300,000 years (3.29-2.97 Ma) of the mid-Pliocene warm period, is not sufficient to cause the deglaciation scenarios proposed by Harwood & Webb (1986) based on Sirius Group diatom deposition. Conversely, even the largest reconstructed extents of the EAIS are smaller than modern, suggesting that the inferences made from Dry Valleys landscape stability cannot be extended to the entire continent. The Pliocene seems to have experienced significant changes in cryosphere volume and extent, with temperature increases within the range predicted for the coming century (IPCC 2001). Ice sheet volumes and extents do not evolve linearly with temperature. Therefore hysteresis in the ice sheet – climate system may mean a similar future temperature increase produces less cryospheric change than seen in the Pliocene. However, there are many examples of potentially climate induced changes in modern ice sheets and palaeoclimate studies of warmer climates, such as the Pliocene, do not seem to rule out significant changes to the cryosphere.

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Tables

Table 1: Pliocene Arctic palaeoenvironmental data set.

Site	Location	Evidence	Paper
Kap København	Greenland	Macrofossils, ostracoda, foraminifera	Funder et al. 2001
Île de France	Greenland	Sedimentology, macrofossils, foraminifera	Bennike et al. 2002
Lodin Elv	Greenland	Ostracoda	Penney 1993
North GRIP	Greenland	Macrofossils (in basal ice)	Dahl-Jensen 2006
Tjornes Peninsula	Iceland	Palynology	Willard 1994
ODP Sites 910 & 911	Arctic Ocean	Palynology	Willard 1996
ODP Site 986	Arctic Ocean	Dinoflagellates	Smelror 1999
ODP Site 907	Norwegian Sea	IRD	Jansen et al. 2000
ODP Site 642	Norwegian Sea	Palynology	Willard 1994
BGS Borehole 88/7	Northern Atlantic	Sedimentology, nanofossils	Stoker et al. 1994
ODP Site 918	Northern Atlantic	IRD	St. John & Krissek 2002
ODP Site 646	Northern Atlantic	Palynology	Willard 1994
ODP Site 645	Baffin Bay	Palynology, IRD	de Vernal & Mudie 1989
Ellesmere Island	Canadian Arctic	Palynology, macrofossils	Csank 2006
Meighen Island	Canadian Arctic	Macrofossils	Matthews 1987
Prince Patrick Island	Canadian Arctic	Macrofossils	Fyles 1990
Northern Banks Island	Canadian Arctic	Macrofossils	Fyles et al. 1994
Bluefish Basin	Yukon, Canada	Palynology, macrofossils	Matthews & Ovenden 1990
Niguanak Site	Alaska	Macrofossils	Matthews & Ovenden 1990
Porcupine River	Alaska	Macrofossils	Matthews & Ovenden 1990
Lost Chicken Mine	Alaska	Palynology, macrofossils	Matthews et al. 2003
Yakataga Formation	Alaska	IRD	Lagoe & Zellers 1996
Alaska Range	Alaska	Palynology	Ager 1994
Kugruk River	Alaska	Palynology	Matthews & Ovenden 1990
Anadyr Basin	Arctic Russia	Palynology, macrofossils	Fradkina 1991
Kolyma River	Arctic Russia	Palynology, macrofossils	Fradkina 1991
Magadan District	Arctic Russia	Palynology, macrofossils	Fradkina 1991

Table 2: Pliocene Antarctic sites.

Site	Location	Evidence	Paper
Cockburn Island	Northern Antarctic Peninsula	Micropalaeontology	Zinsmeister & Webb 1988
James Ross Island	Northern Antarctic Peninsula	Micropalaeontology, macrofossils	Jonkers et al. 2002
Seal Nunataks	Northern Antarctic Peninsula	Volcanology	Smellie & Hole 1997
Hornpipe Heights	Alexander Island	Volcanology	Smellie 1999
ODP Site 1095	Western Antarctic Peninsula	Sedimentology, geochemistry	Hepp et al. 2006
ODP Site 1096	Western Antarctic Peninsula	Sedimentology	Hillenbrand & Ehrmann 2005
ODP Site 1097	Western Antarctic Peninsula	Sedimentology	Hillenbrand & Ehrmann 2005
ODP Site 1101	Western Antarctic Peninsula	Ice rafted debris	Cowan 2002
Mount Sidley	West Antarctic Ice Sheet	Volcanology	Panter et al. 1994
Mount Murphy	West Antarctic Ice Sheet	Diatoms	LeMasurier et al. 1994
Crary Ice Rise	Ross Sea Ice Streams	Diatoms	Scherer et al. 1988
Upper Whillans	Ross Sea Ice Streams	Diatoms	Scherer 1991
ODP Site 693	Weddell Sea	Sedimentology	Barker & Kennett 1988
ODP Site 694	Weddell Sea	Sedimentology	Barker & Kennett 1988
Bardin Bluffs	Amery Basin	Diatoms	Whitehead et al. 2004
Sørsdal Formation	Amery Basin	Diatoms	Whitehead et al. 2001
Oliver Bluffs	Transantarctic Mountains	Palaeontology, sedimentology	Francis & Hill 1996
Dry Valleys	Transantarctic Mountains	Geomorphology	Sugden et al. 1995
DVDP Holes 10 & 11	Transantarctic Mountains	Micropalaeontology	Ishman & Reick 1992
ODP Site 1165	Prydz Bay	Diatoms	Whitehead et al. 2005
ODP Site 1166	Prydz Bay	Diatoms	Whitehead et al. 2005
Gunnerus Ridge	Enderby Land	Sedimentology	Hillenbrand & Ehrmann 2003
DSDP Site 274	Transantarctic Mountains	Palynology	Fleming & Barron 1996

Figure Captions

Figure 1: Comparison of temperature data sets for the (**A**) Cenozoic (Lear *et al.* 2000), (**B**) last four glacial cycles (Petit *et al.* 1999) and (**C**) last thousand years (Mann *et al.* 1999) with predictions of future global temperatures.

Figure 2: Pliocene palaeoenvironmental data for the Arctic (Table 1), superimposed on the 1°x1° modern vegetation distribution (Matthews, 1983).

Figure 3: Antarctic Pliocene sites (Table 2) and Sirius Group deposits (Stroeven, 1997).

Figure 4: Differing views of Cenozoic glaciations (**a**) periods of proposed Antarctic deglaciation, dated from Sirius Group diatoms (Harwood, 1983) (**b**) glaciations (solid represent permanent and dashed transient) inferred from Cenozoic $\delta^{18}O$ values (Zachos *et al.* 2001).

Figure 5: Mid-Pliocene temperatures, utilised as boundary conditions within ice sheet models, as simulated by HadAM3 GCM (Haywood *et al.* 2000). (**A**) Difference between mid-Pliocene and modern annual temperatures (ºC). (**B**) Mid-Pliocene Antarctic January temperatures (ºC).

Figure 6: (**A**) Isostatically rebounded Antarctic bed topography (Lythe *et al.* 2001). (**B**) Basin mask applied to Antarctic bed topography in order to change the depth of subglacial basins. Quoted change is multiplied by this field, enabling smooth transition zones.

Figure 7: Largest (**1**) and smallest (**2**) BASISM reconstructions of the mid-Pliocene EAIS under the modelled Hadley Centre GCM climate. (**a**) Surface height, (**b**) ice velocities and (**c**) mass balance of the modelled ice sheets.

Figure 8: Results of a sensitivity study of modelled temperatures and basin altitude for the EAIS. For each scenario the depicted ice sheet represents the maximum

marine incursion into the subglacial basins and minimum ice sheet extent. These reconstructions are not representative of the equilibrium state, but also capture any transient ice sheet responses.

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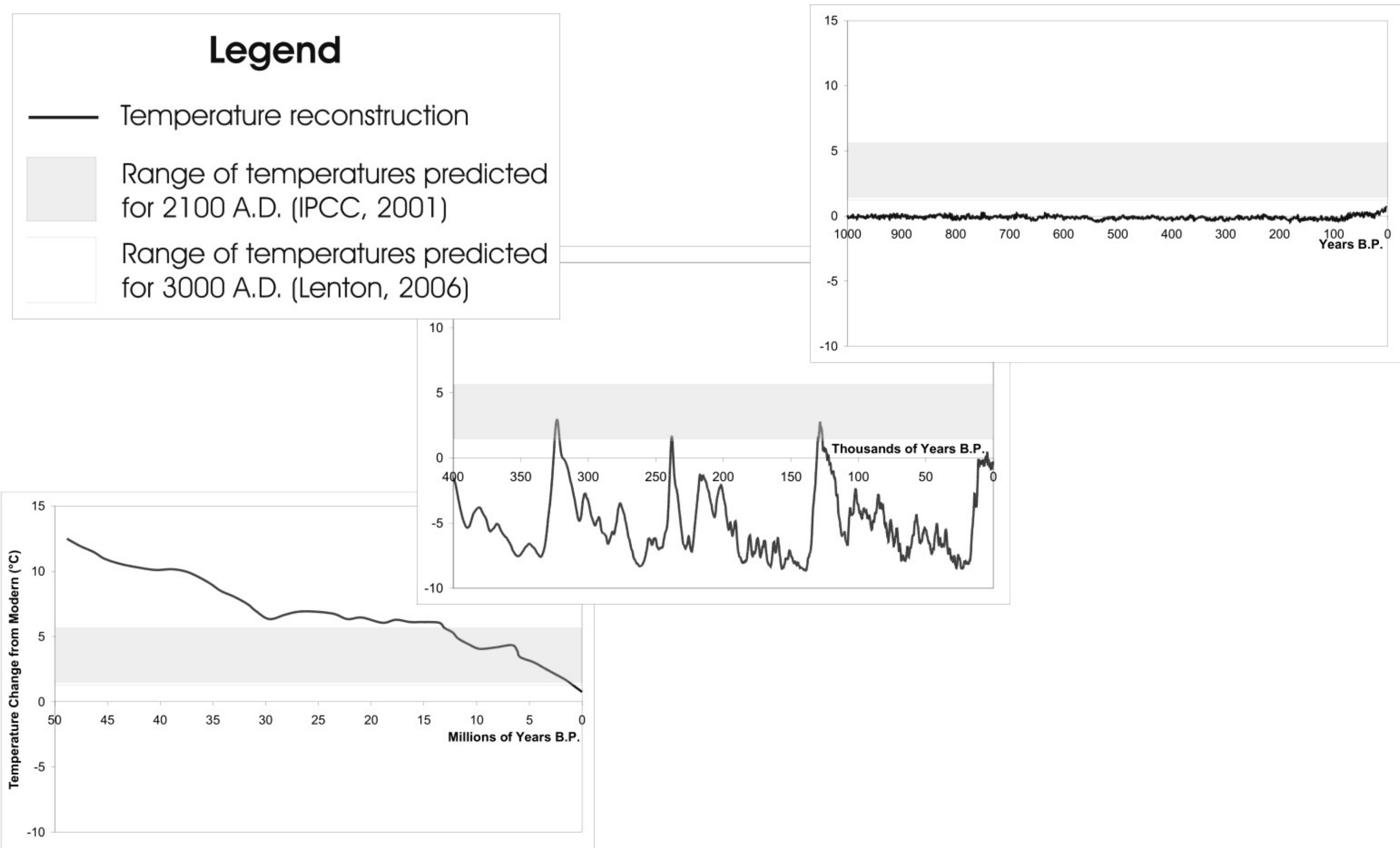
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Temperature reconstruction
Range of temperatures predicted for 2100 A.D. (IPCC, 2001)
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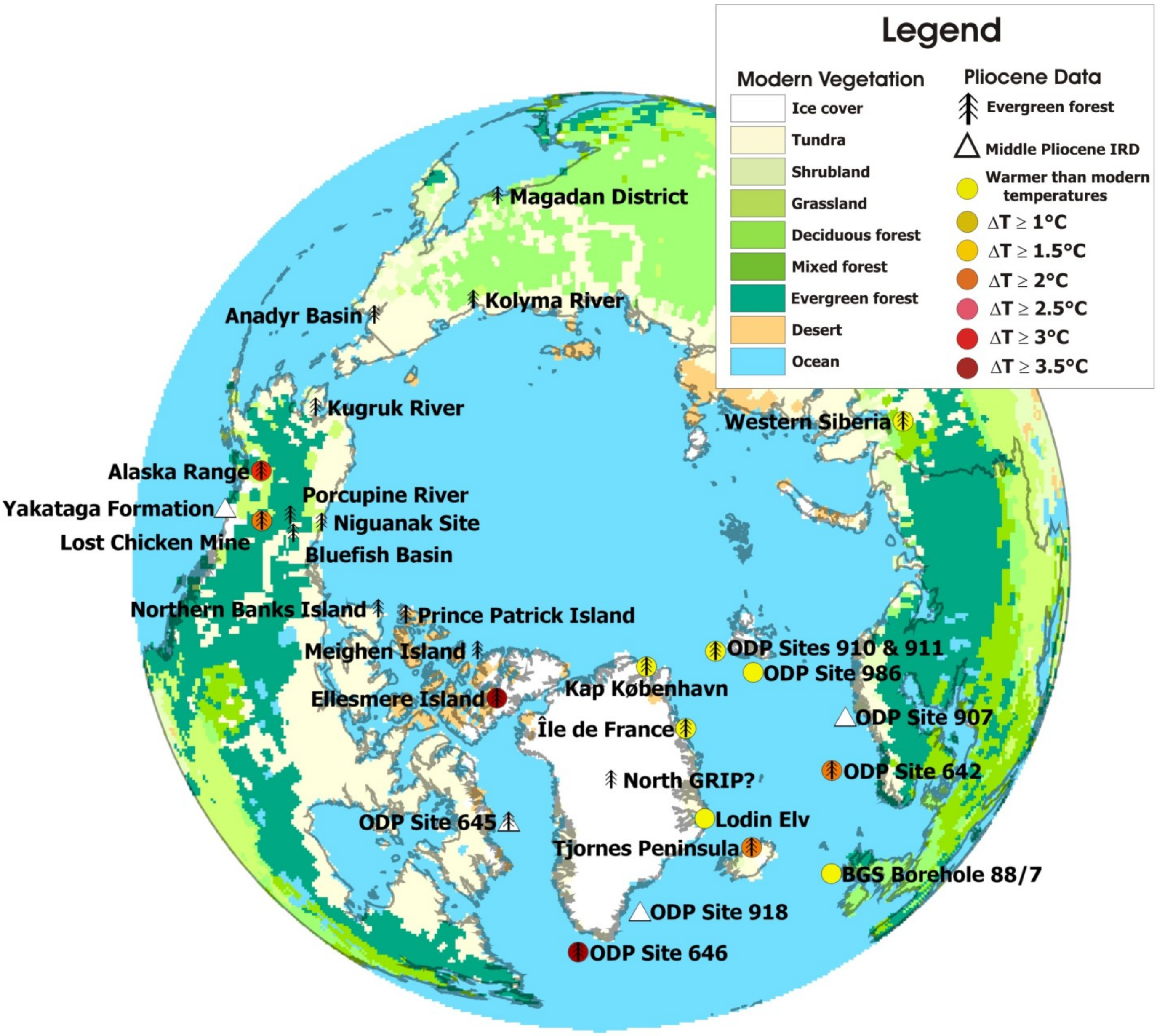
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Middle Pliocene IRD
Warmer than modern temperatures
ΔT ≥ 1°C
ΔT ≥ 1.5°C
ΔT ≥ 2°C
ΔT ≥ 2.5°C
ΔT ≥ 3°C
ΔT ≥ 3.5°C
Magadan District
Kolyma River
Anadyr Basin
Kugruk River
Western Siberia
Alaska Range
Yakataga Formation
Porcupine River
Niguanak Site
Lost Chicken Mine
Bluefish Basin
Northern Banks Island
Prince Patrick Island
Meighen Island
ODP Sites 910 & 911
ODP Site 986
Ellesmere Island
Kap København
ODP Site 907
Île de France
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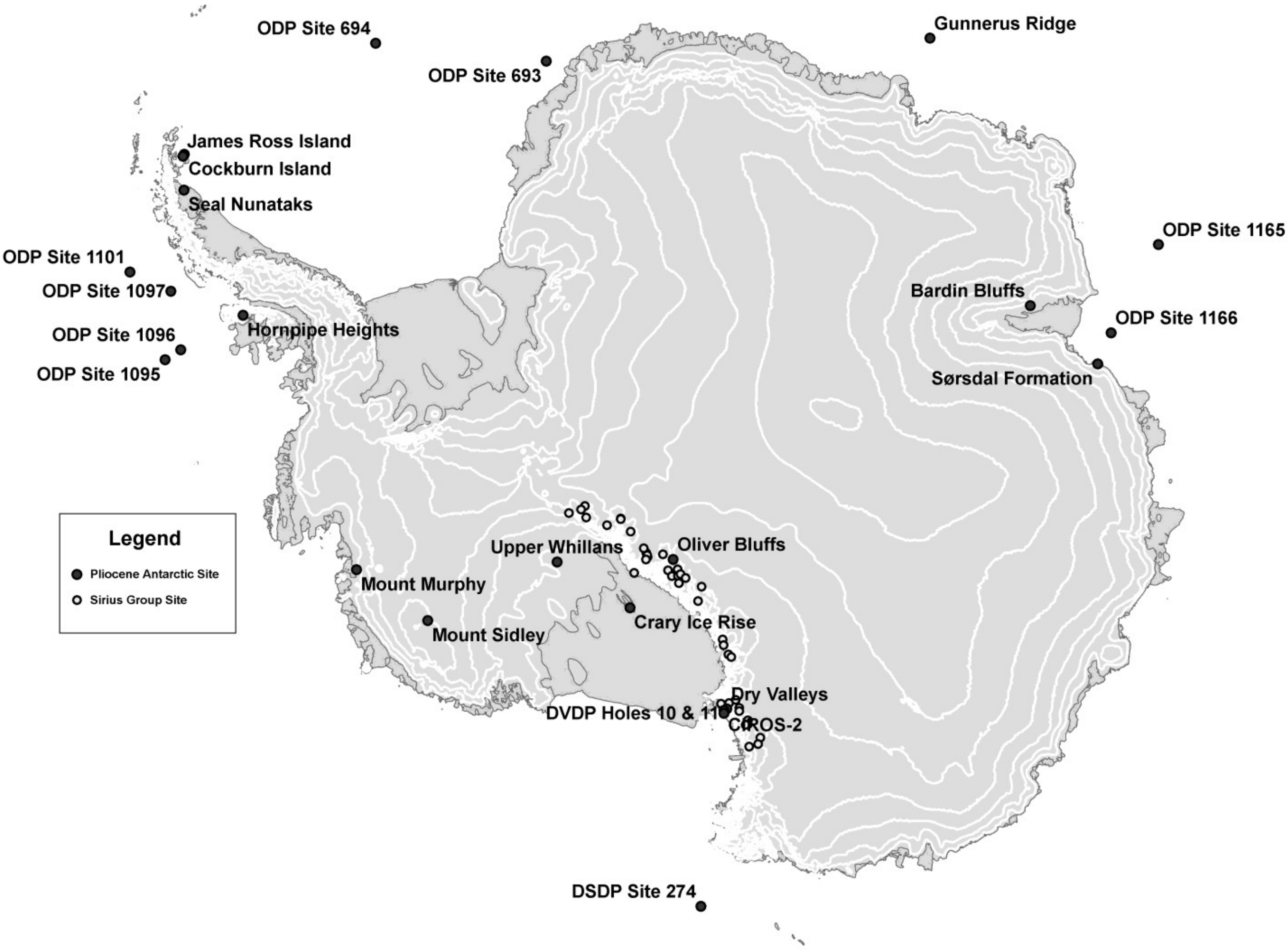
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ODP Site 1166
Hornpipe Heights
ODP Site 1096
ODP Site 1095
Sørsdal Formation
Legend
Pliocene Antarctic Site
Sirius Group Site
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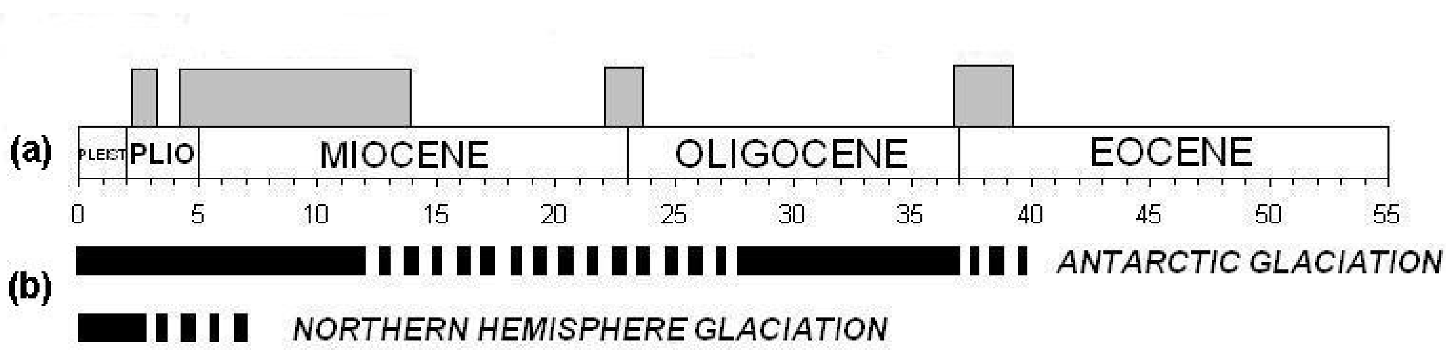
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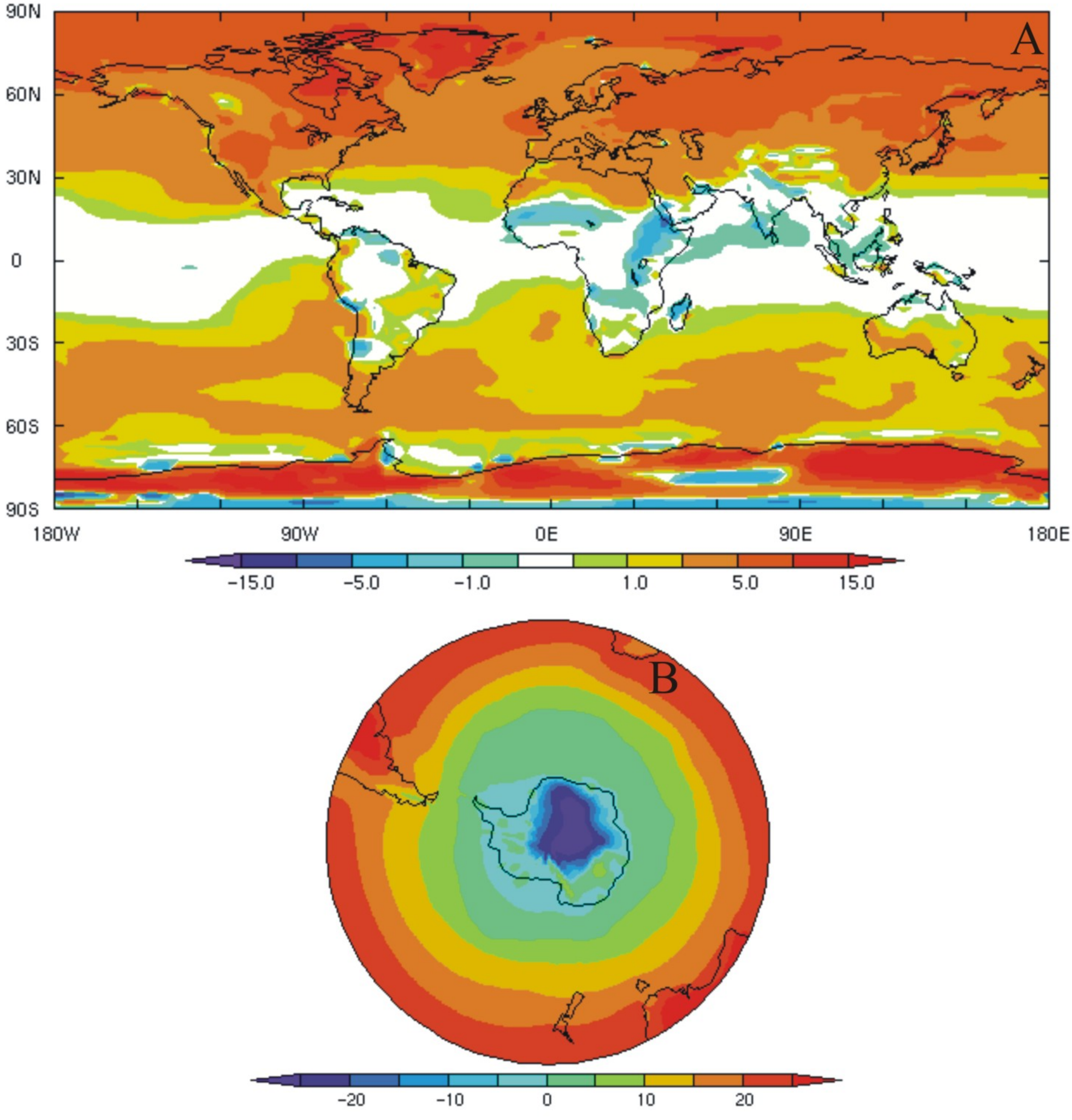
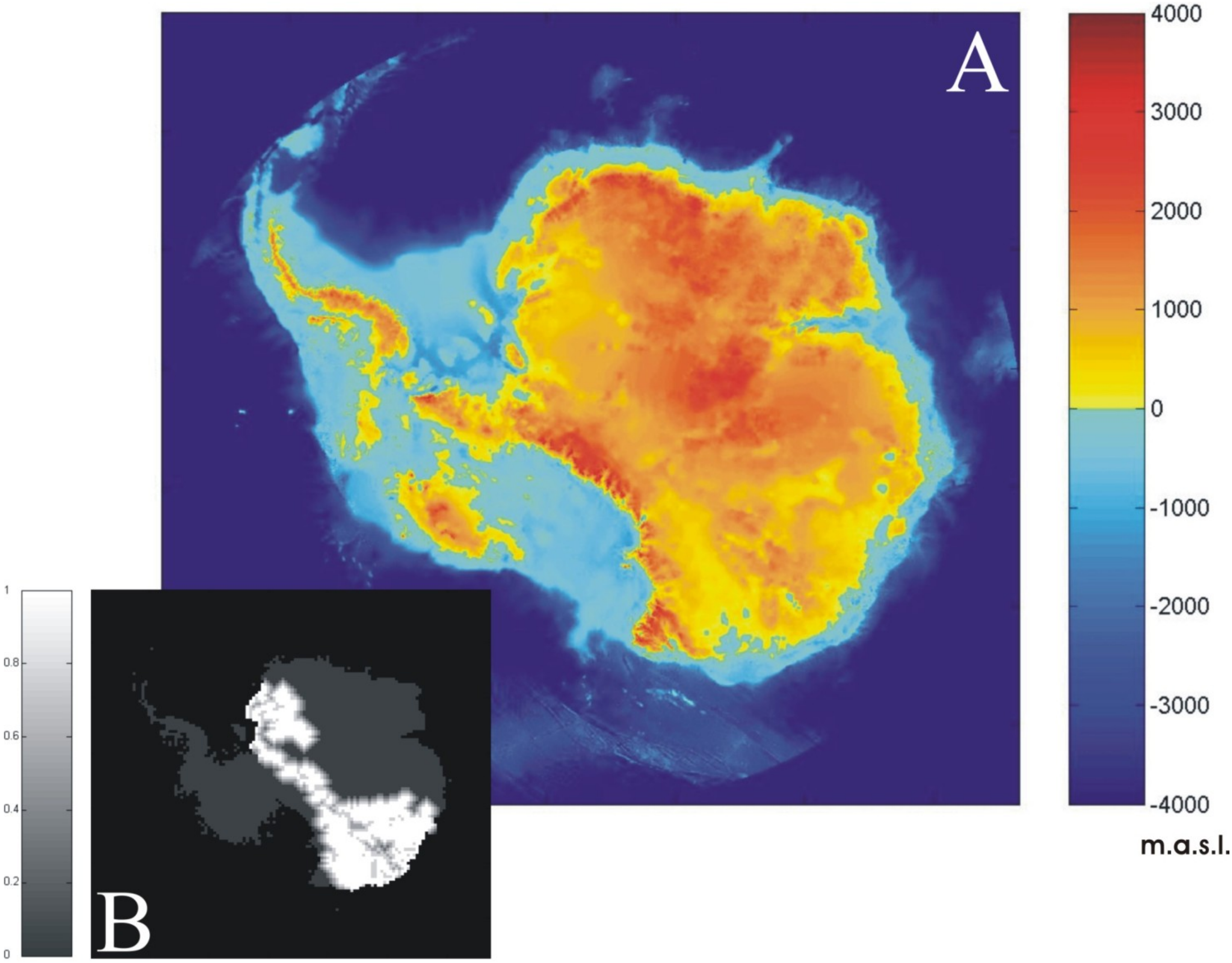
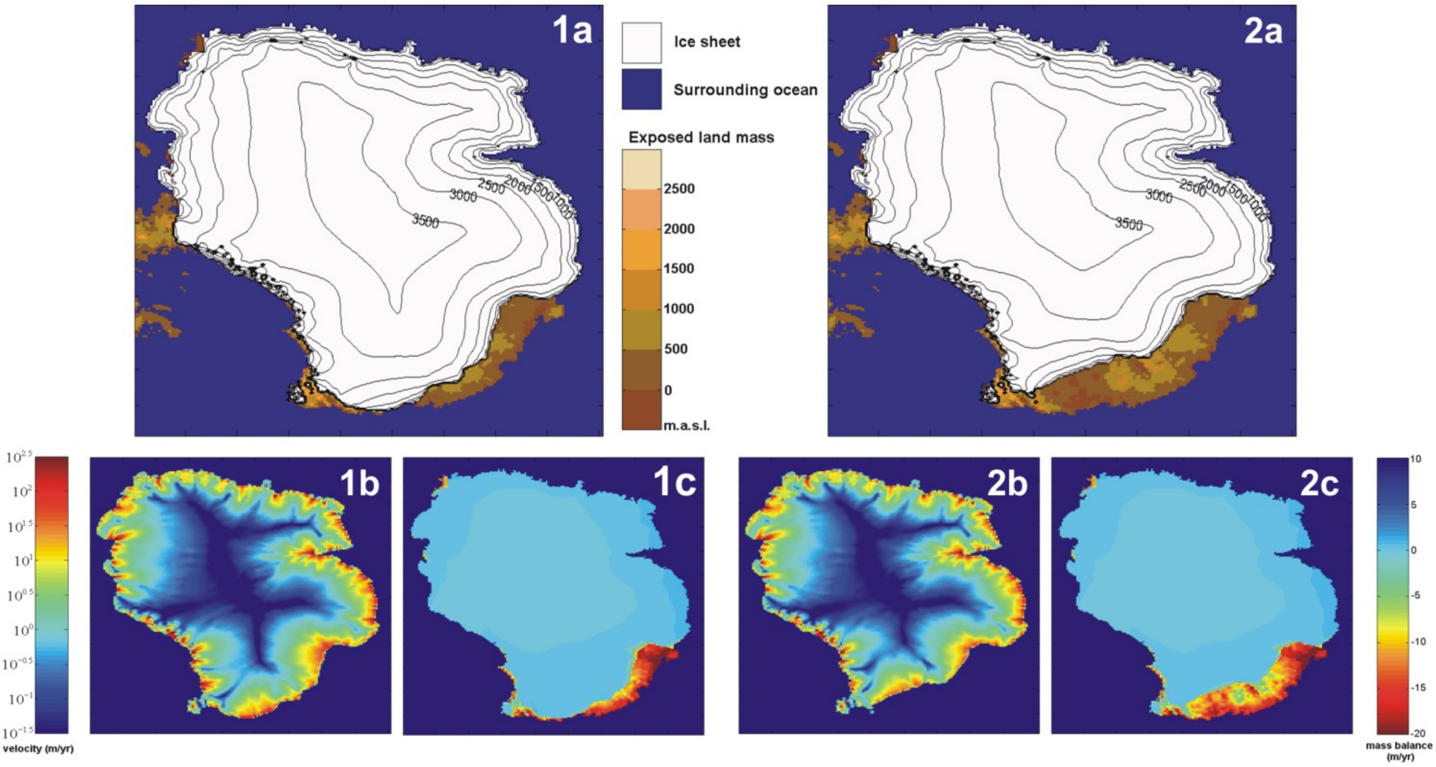
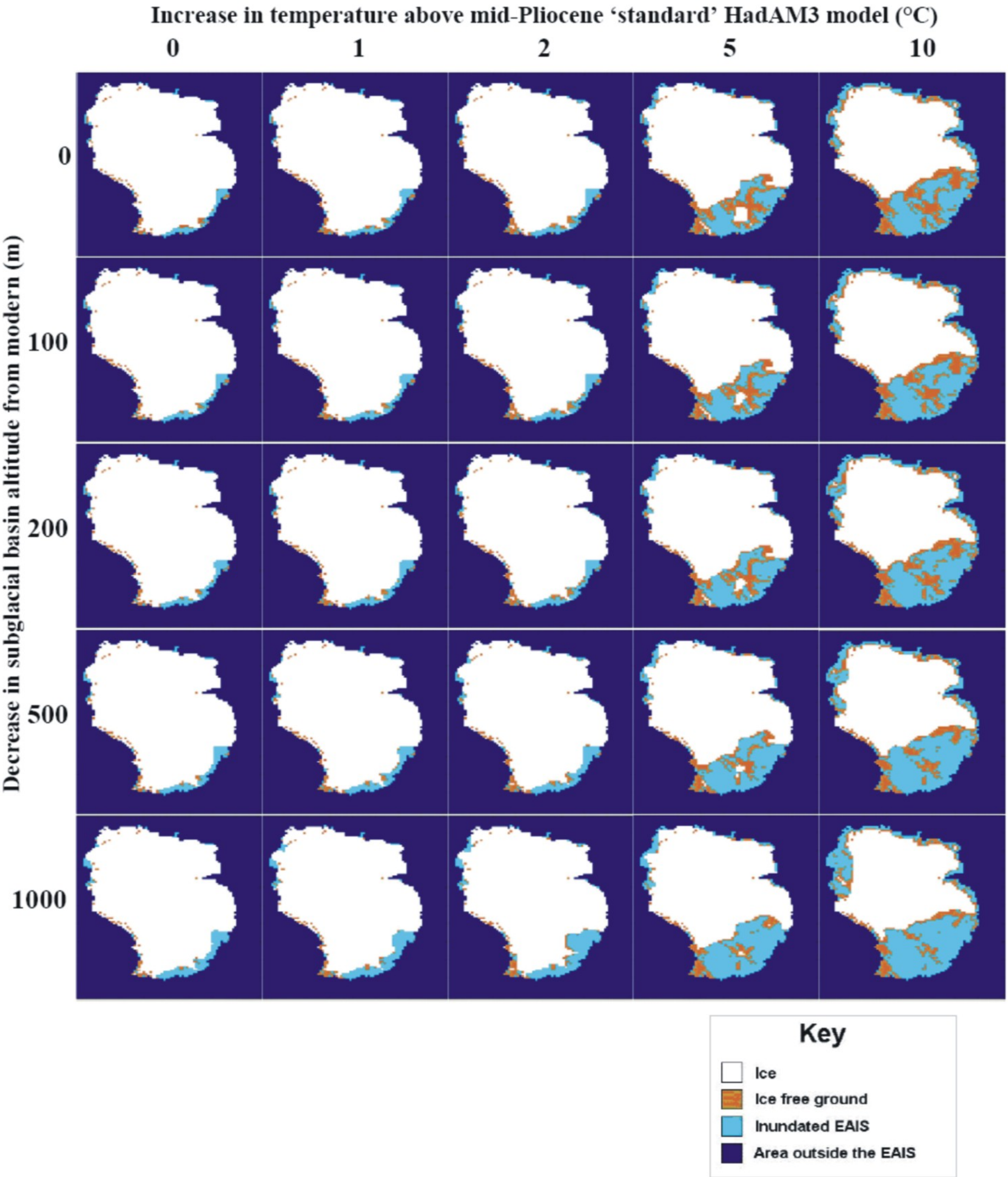
0
5
10
15
20
25
30
35
40
45
50
55
(b)
ANTARCTIC GLACIATION
NORTHERN HEMISPHERE GLACIATION


90N
60N
30N
0
30S
60S
90S
180W
90W
0E
90E
180E
A
-15.0
-5.0
-1.0
1.0
5.0
15.0
B
-20
-10
0
10
20


A
4000
3000
2000
1000
0
-1000
-2000
-3000
-4000
m.a.s.l.
B
1
0.8
0.6
0.4
0.2
0


1a
2a
Ice sheet
Surrounding ocean
Exposed land mass
2500
2000
1500
1000
500
0
m.a.s.l.
3500
3000
2500
2000
1500
1000
1b
1c
2b
2c
velocity (m/yr)
mass balance (m/yr)
10
5
0
-5
-10
-15
-20


Increase in temperature above mid-Pliocene 'standard' HadAM3 model (°C)
0
1
2
5
10
Decrease in subglacial basin altitude from modern (m)
0
100
200
500
1000
Key
Ice
Ice free ground
Inundated EAIS
Area outside the EAIS