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Chapter 9

The Oligocene–Miocene Boundary – Antarctic Climate Response to Orbital Forcing

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ABSTRACT

Recent high-resolution Oligocene–Miocene oxygen isotopic records revealed a relatively transient, ca. *2 myr period, 1‰ amplitude cyclicity in isotopic values (Oi and Mi events, respectively). Intriguingly, it has been suggested that these isotopic excursions in oceanic $\delta^{18}O$ were linked to ephemeral growth and decay in Antarctic ice sheets. A great deal of effort in the palaeoceanography community has been focused on developing techniques and gathering further records to determine if the Antarctic ice has behaved in such a transient manner in the past and indeed what factors might have led to the rapid growth and decay of ice sheets. Deciphering between temperature and ice-volume influences in the deep-sea isotopic record has proven somewhat difficult. Approaches have included the sampling of sediment from beneath different water masses, development of an independent palaeothermometer using magnesium/calcium ratios and improving the resolution and accuracy of coastal sea-level records. Despite these advances it* AU :2

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is only through the recovery of Antarctic drill core records that we have been able to test the resulting hypotheses. Combined with numerical climate models, ice-volume estimates are also available.The Antarctic Oligocene–Miocene record is most complete in the Victoria Land Basin as recovered in the CIROS-1 and CRP-2A drill ~~holes~~. The strata recovered in both drill ~~holes~~ are cyclic in nature and interpreted to represent periodic advance and retreat of ice across the Antarctic margin concomitant with sea-level fall and rise, respectively. While, environmental data suggest a significant Antarctic climate threshold across the Oligocene–Miocene boundary with cooler temperatures implied in the early Miocene, ice volume and palaeo-sea-level estimates suggest a significant but transient growth in the Antarctic ice sheet to ~25% larger than present. The Antarctic data are entirely consistent with the predictions from deep-sea records including the suggestion that the glacial advance was relatively short lived and interglacial conditions were re-established within a few hundred thousand years.The duration and transience of the Mi1 glacial expansion and swift recovery in Antarctica likely resulted from the limited polar summer warmth from the coincidence of low eccentricity and low-amplitude variability in obliquity of the earth's orbit at the Oligocene–Miocene boundary. This was followed by warmer polar summers and increased melt from increased eccentricity and high-amplitude variability in obliquity in the early Miocene, allowing the recovery of vegetation on the craton. Atmospheric CO_2 concentrations remained below a 2× pre-industrial threshold, which promoted sensitivity of the climate system to orbital forcing. While climate and ice-sheet modelling support the fundamental role of greenhouse gas forcing as a likely cause of events like Mi1, the models underestimate the range of orbitally paced ice-sheet variability recognised in early Miocene isotope and sea-level records unless accompanied by significant fluctuations in greenhouse gas concentrations. While tectonic influence may have been secondary, they may well have contributed to oceanic cooling recorded at the Cape Roberts Project site in the South Western Ross Sea.

9.1. Introduction

The paucity of Cenozoic outcrop on the Antarctic craton has led to the reliance on proxy records (isotopic signatures in microfossils, deep-sea erosion events and former sea levels on distal continental margins) to help unravel the history of climate and ice sheets on Antarctica (Kennett and Shackleton, 1976; Kennett, 1977; Wright and Miller, 1993; Miller and Mountain, 1996; Miller et al., 2005; Zachos et al., 2001a). Much attention has been focused ~~in~~ the search for Antarctic data sets to 'ground truth' significant climate trends, events and thresholds observed in these proxy records. Drilling around the Antarctic margin by seven legs of the Deep Sea Drilling Project (DSDP; Kennett et al., 1974; Hayes and Frakes, 1975) and Ocean Drilling Programme (ODP; Barker et al., 1988a, b; Ciesielski et al., 1988; Barron et al., 1989; Barker et al., 1999; O'Brien et al., 2001) and several sea-ice-based drilling projects (Barrett, 1986, 1989; Cape Roberts Science Team, 1998, 1999, 2000; Fig. 9.1) has recovered Cenozoic sequences, which have allowed the testing of interpretations of Antarctic Glacial history from proxy records and climate models. While the DSDP and ODP core recovery has been between 14 and 40%, riser drilling from sea ice in the South Western Ross Sea has enabled recovered of some high-quality intervals of the Cenozoic (95–98%) recovering *prima facie* documentation of climate and cryospheric changes in Antarctica. One interval that is particularly well-sampled and well-dated in several drill cores is the Oligocene–Miocene boundary, permitting an accurate comparison to deep-sea high-resolution isotopic records from lower latitudes ~~has been possible~~ (Naish et al., 2001, 2008; Wilson et al., 2002; Roberts et al., 2003).

This chapter reviews recent evidence for a glacial expansion in Antarctica coincident with the Oligocene–Miocene boundary and the Mi1 deep-sea oxygen isotope excursion. The climatic significance of the boundary has only recently become apparent from recalibration of the Oligocene–Miocene time scale using astrochronology (Zachos et al., 2001b). Consequently, age data and chronstratigraphy of the Oligocene–Miocene boundary and the Antarctic strata that contain the boundary are also reviewed. Data sets considered to indicate climate and ice-sheet variability across the ~~B~~oundary include benthic and planktic oxygen isotope ($\delta^{18}O$) records (Kennett and Shackleton, 1976; Miller et al., 1991; Wright and Miller, 1992; Paul et al., 2000; Zachos et al., 2001b; Billups et al., 2002) and microfossil geochemistry (Billups and Schrag, 2002; Lear et al., 2004), sequence stratigraphic analyses of Antarctic (Fielding et al., 1997; Naish et al., 2001, 2008) and mid-latitude (Kominz and Pekar, 2001; Pekar et al., 2002; Miller et al., 2005~~, 2006~~; Pekar and DeConto, 2006) AU :3 continental margin strata, Antarctic palaeobotany and palynology (Askin and Raine, 2000; Barrett, 2007) and physical properties of Antarctic drill core strata including lithology, clay mineralogy, mudrock geochemistry and magnetic mineralogy (Verosub et al., 2000; Ehrmann et al., 2005; Passchier and Krissek, 2008). Finally, the cause of the Mi1 glaciation is considered.

9.1.1. Identification of the Oligocene–Miocene Boundary

Early definition of the Oligocene–Miocene boundary relied on the identification of the last occurrence of the calcareous nannofossil *Dictyococcites*

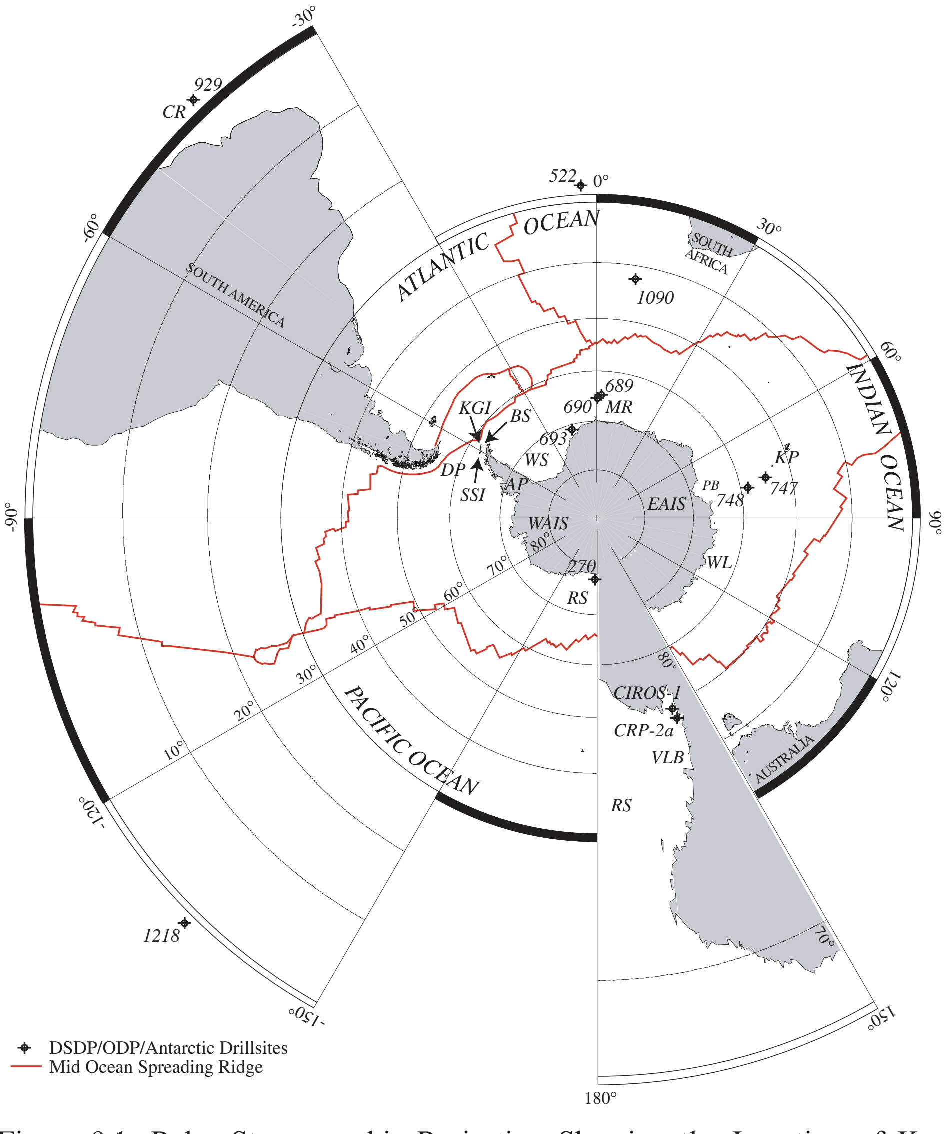

Figure 9.1: Polar Stereographic Projection Showing the Location of Key Oligocene–Miocene Boundary Sections Discussed in the Text. Abbreviations of Locations: Numbers are DSDP/ODP Drill sites; AP, Antarctic Peninsula; BS, Bransfield Strait; CIROS, Cenozoic Investigations in the Ross Sea Drilling Project; CR, Ceara Rise; CRP, Cape Roberts Drilling Project; DP, Drake Passage; EAIS, East Antarctic Ice Sheet; KGI, King George Island; KP, Kerguelen Plateau; MR, Maud Rise; PB, Prydz Bay; RS, Ross Sea; SSI, South Shetland Islands; VLB, Victoria Land Basin; WAIS, West Antarctic Ice Sheet; WS, Weddell Sea.

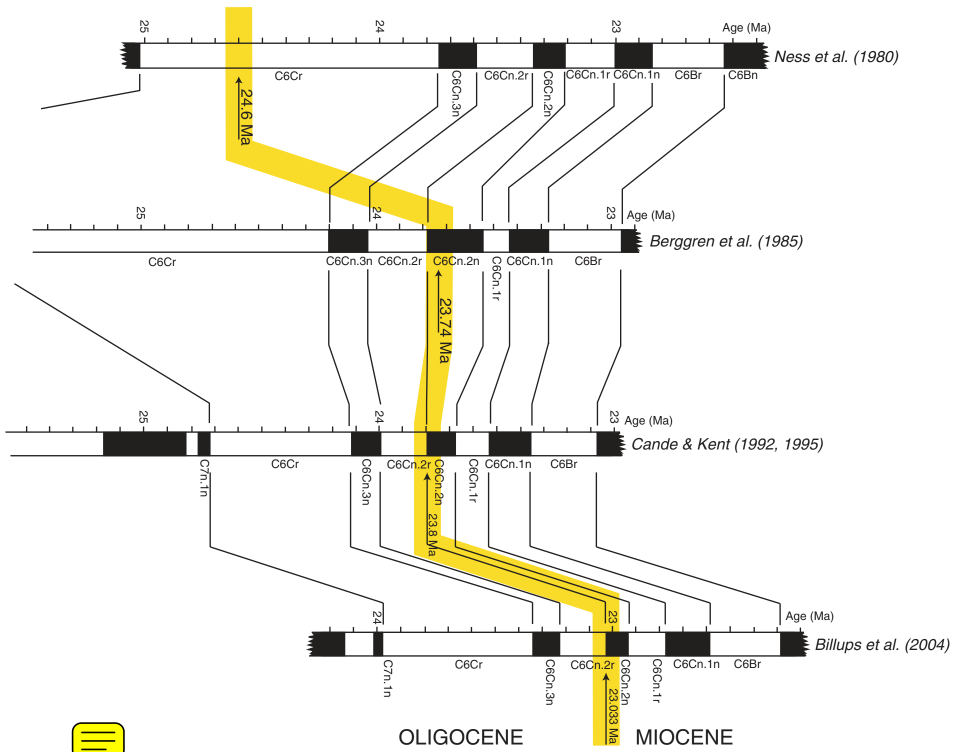

Figure 9.2: History of the Calibration of the Geomagnetic Polarity Time Scale in the Vicinity of the Oligocene–Miocene Boundary. Boundary Position in Ness et al. (1980) and Berggren et al. (1985) is Fixed to the Last Occurrence of the Nannofossil *Reticulofenestra bisectus* and in Cande and Kent (1992, 1995), Gradstein et al. (2004) and Billups et al. (2004) is Fixed to the Base of Magnetic Polarity Subchron C6Cn.2n. Ness et al. (1980), Berggren et al. (1985), Cande and Kent (1992, 1995) are Calibrated by Various Mid-Ocean Ridge Spreading Rate Models. Billups et al. (2004) and Gradstein et al. (2004) are Calibrated by Astronomical Tuning at ODP Site 1090.

bisectus (23.7 Ma; Berggren et al, 1985). However, this has proved problematic in the colder waters, coarser sediments and hiatus prone strata of the Antarctic and Southern Ocean. The reassignment of the boundary by Cande and Kent (1992, 1995) to the slightly older base of Magnetic Polarity subchron C6Cn.2n (23.8 Ma; Fig. 9.2) has made its identification more straightforward in Antarctica and the Southern Ocean but only in relatively complete and continuous stratigraphic successions (Wilson et al., 2002; Roberts et al., 2003). More recently, the recognition of astronomically influenced cyclical physical properties and $\delta^{18}O$ records in continuously deposited deep successions has enabled astronomical calibration of late Oligocene through early Miocene time. The astronomical calibration suggested that, while still coincident with the base of subchron C6Cn.2n, the boundary was in fact nearly a million years younger (22.9 ± 0.1 Ma; Shackleton et al., 2000; Pälike et al., 2004; 23.03 Ma; Billups et al., 2004; Gradstein et al., 2004; Fig. 9.2). The climatic significance of this was outlined by Zachos et al. (2001b) and Pälike et al. (2006) who recognised the coincidence of the Oligocene–Miocene boundary and the Mi1 isotope excursion with an unusual coincidence of low eccentricity and low-amplitude variability in obliquity of the earth's orbit (Fig. 9.3). This would have placed the earth in a sustained period of unusually low seasonality (cold summers), which Zachos et al. (2001b) claimed would have limited polar summer warmth and encouraged ice growth at the poles. Equally, within a few hundred thousand years, the coincidence of increased eccentricity and high-amplitude variability in obliquity would have resulted in warmer polar summers and increased summer melt.

9.2. Proxy Records

9.2.1. The Isotopic Record

Oxygen isotope ratios ($\delta^{18}O$) in foraminiferal tests from deep-sea sedimentary records have long been recognised to represent the Cenozoic climatic (temperature, sea level and ice volume) history of the earth (e.g. Shackleton et al., 1977 and references therein). However, deciphering the climatic history of Antarctica from $\delta^{18}O$ values alone in deep-sea records has always proven difficult due to the ambiguity of influence on the signal from the volume of ice on land versus isotopic fractionation, which is related to the water temperature during the precipitation of calcite (Miller et al., 1991). Early studies attempted to separate the two influences by focusing their analyses on foraminiferal species, such as benthic forms, know to live in water masses thought to be relatively stable in their temperature history (Shackleton and Kennett, 1975; Kennett, 1977), hence deducing that any shorter-term fluctuation was due to ice volume rather than temperature fractionation. While these studies recognised major threshold changes in the climatic deterioration of Antarctica, assuming no Northern Hemisphere ice sheets at the time, their resolution was limited and the record incomplete across the Oligocene–Miocene boundary. In a higher-resolution compilation of

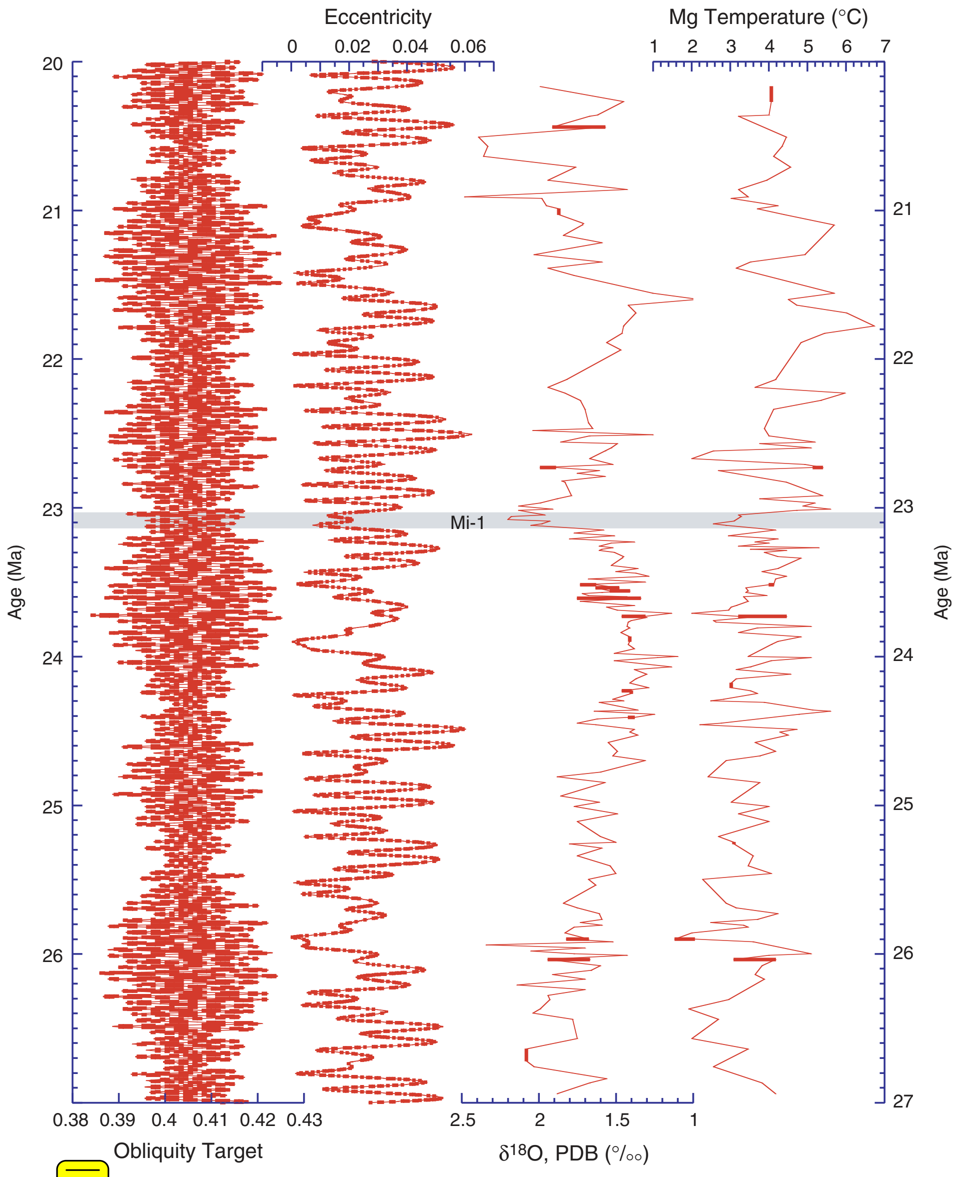

Figure 9.3: Oxygen Isotope and Magnesium Temperature Data Across the Oligocene–Miocene Boundary Interval (Data from Lear et al., 2004). AU :1 Obliquity and Eccentricity Orbital Target Data from Laskar et al. (2003). All Data are Plotted Against the Astronomical Time Scale Presented by Billups et al. (2004).

Oligocene–Miocene $\delta^{18}O$ records, Miller et al. (1991) recognised a relatively transient *ca.* 2 myr period and 1‰ amplitude cyclicity in Miocene isotope values (isotope events Mi1-Mi7). The 1‰ shifts, they suggested were of the same order as the threshold shifts identified by earlier studies (Kennett and Shackleton, 1976) and represent similar volumes of ice accumulation on the Antarctic craton. The most significant of these shorter order isotopic events, Mi1, was coincident with the Oligocene–Miocene boundary (Fig. 9.3). Originally defined by Miller et al. (1991) from DSDP Site 522 (Figs. 9.1 and 9.3), the event has subsequently been confirmed at numerous locations and the timing, duration and magnitude refined (Zachos et al., 1997; Paul et al., AU :4
2000; Zachos et al., 2001a, b; Billups et al., 2002, 2004, Billups and Schrag, 2002).

For at least 2 myr prior to the Oligocene–Miocene boundary, $\delta^{18}O$ values were relatively stable and of low-amplitude variability (<0.5‰; Paul et al., 2000). In contrast, the Mi1 event represents a dramatic $\sim$1‰ increase in $\delta^{18}O$, over a 250 ky period immediately prior to the Oligocene–Miocene boundary peaking coincident with the boundary (Billups et al., 2004; Fig. 9.3). Peak values persisted for only $\sim$20 ky before returning to similarly low amplitude but slightly increased late Oligocene mean $\delta^{18}O$ values over the first $\sim$120 ky of the early Miocene (Paul et al., 2000). The covariance of the isotope signal in both benthic and planktic species in the late Oligocene at equatorial Atlantic ODP Site 929 led Paul et al. (2000) to conclude that the variability was primarily ice volume driven. However, the Mi1 event, itself, is of relatively lower amplitude in planktic records which led Paul et al. (2000) to suggest that only 0.5‰ is likely due to ice-volume effects, which, using the late Pleistocene calibration, represents growth of an ice sheet in Antarctica of similar proportion to the present-day East Antarctic ice sheet (EAIS). The remaining 0.5‰, they concluded was due to a 2°C cooling of bottom waters at ODP Site 929 in the western Equatorial Atlantic.

Another approach to deciphering temperature versus ice-volume components of the deep-sea $\delta^{18}O$ signal was employed by Lear et al. (2004) who determined palaeotemperature independently from Mg/Ca ratios in foraminifera tests across the Oligocene–Miocene boundary at ODP Site 1218 in the eastern Equatorial Pacific (Figs. 9.1 and 9.3). The 2–3°C of cooling predicted in the equatorial Atlantic by Paul et al. (2000) immediately prior to Oligocene–Miocene boundary was confirmed in the equatorial Pacific by Lear et al. (2004). However, peak cooling preceded peak $\delta^{18}O$ values at ODP Site 1218 and a slight (1°C) warming in bottom waters of the equatorial pacific is recorded immediately prior to and during peak Mi1 $\delta^{18}O$ values followed by another 2°C of warming post-Mi1. A similar temperature variation was observed $\sim$800 ky prior to the Oligocene–Miocene boundary.

Lear et al. (2004) concluded that the Mi1 event did indeed represent a significant increase in continental ice volume and, given the warming influence predicted by Mg/Ca ratios, perhaps a larger ice-volume growth than predicted by Paul et al. (2000). Billups and Schrag (2002) also suggested that the $\delta^{18}O$ record from ODP Site 747 represented an ice-volume signal because paired Mg/Ca measurements suggested little change in ocean temperature through the early Miocene. Recent work, however, suggests caution when interpreting stable intervals in Mg/Ca ratios from deep-water sites due to potential saturation of carbonate, which might affect the partitioning of Magnesium into benthic foraminifera (Elderfield et al., 2006; Lear et al., 2008).

9.2.2. Palaeo Sea Levels

Sequence and seismic stratigraphy has provided a means of relating the geologic record of continental margins to global sea-level changes that are often related to ice-volume changes at high latitudes (Vail et al., 1977; Haq et al., 1987). Sea-level history is deduced by the recognition of unconformity-bounded units (i.e. depositional sequences) deposited in response to a cycle of falling and rising sea level. However, determining the relative role of tectonic subsidence and uplift versus rising and falling global sea-level and hence ice-volume fluctuations (glacioeustasy) is still unresolved, particularly in pre-Pleistocene records (e.g. Macdonald, 1991). Vail et al. (1977) and Haq et al. (1987) attempted to extract the glacioeustatic signal from the comparison of records from several continental margins (Fig. 9.4). However, the large amplitudes of sea level/ice volume, the limited resolution of the resulting record and the use of proprietary data to create the sea-level records, has spurred the scientific community to collect independent data to test the records of Vail et al. (1977) and Haq et al. (1987). The Oligocene–Miocene interval of these records has sparked particular interest. Both Vail et al. (1977) and Haq et al. (1987) predicted sea-level rises and falls of between ~50 and 100 m in the late Oligocene. Vail et al. (1977), however, predicted a fall of some 60 m across the Oligocene–Miocene boundary, whereas Haq et al. (1987) predicted a rise of ~100 m across the Oligocene–Miocene boundary. The proprietary nature of much of the data has precluded resolving this conundrum from the same data set.

An alternative passive margin stratigraphic data set is available from the New Jersey/New York Bight region of North America. Sea-level changes predicted from sequence stratigraphic analysis have recently been calibrated from coring as part of the Ocean Drilling Programme Legs 150X and 174AX

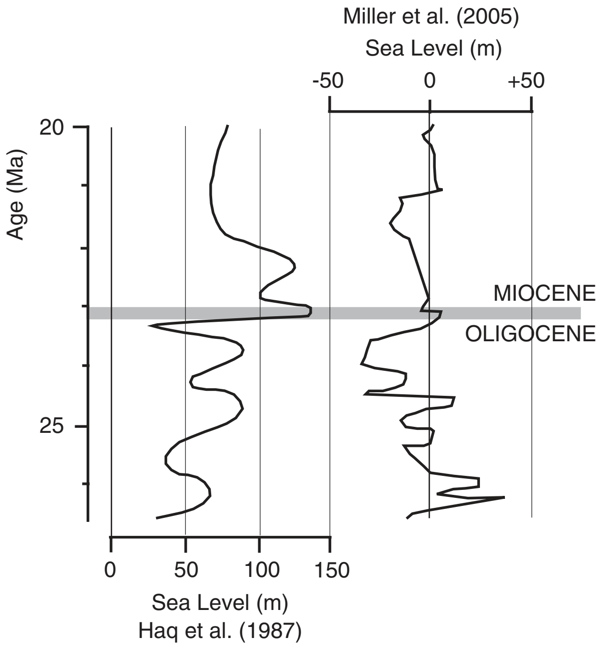

Figure 9.4: Eustatic Sea-Level Curves Derived from Coastal Onlap Patterns for the Oligocene–Miocene Boundary Interval from Haq et al. (1987) and Miller et al. (2005). Calibration of Miller et al. (2005) Curve is Considered More Realistic. Curves are Adjusted to Astronomical Time Scale of Billups et al. (2004).

(Miller and Mountain, 1996; Miller et al., 1997a, b). The glacioeustatic AU :5 contribution to sea-level changes in the Oligocene and earliest Miocene was estimated by combining two-dimensional palaeoslope modelling of the foraminiferal biofacies and lithofacies with two-dimensional flexural back-stripping of the margin (Kominz and Pekar, 2001; Pekar and Kominz, 2001). The depth ranges of foraminiferal biofacies were determined from a combination of standard factor analysis techniques and the backstripped geometries. The geometry of the margin through time was determined using two-dimensional flexural backstripping. Foraminiferal biofacies and lithofacies were then used to constrain the depths of the Oligocene margin profiles obtained from backstripping. A eustatic fall of $\sim 40 \pm 15$ or 56 m apparent sea level (apparent sea level is eustasy plus the effects of water loading on the margin, Pekar et al., 2002) was estimated across the Oligocene–Miocene boundary (Fig. 9.4), which is similar to the glacioeustasy predictions from oxygen isotopic and trace metal geochemical data (Paul et al., 2000; Lear et al., 2004).

Pekar et al. (2006) provided estimates of Antarctic ice volume and the resulting changes in global sea-level for the late Oligocene by applying δ^{18}O-to-sea-level calibrations to deep-sea δ^{18}O records from a number of ODP Sites. Their results indicate that the size of the Antarctic ice sheet increased from approximately 50% of the present-day EAIS during the latest Oligocene to as much as 25% larger than the present-day EAIS at the Oligocene–Miocene boundary. Ice volume returned to near late Oligocene size in the early Miocene (Pekar and DeConto, 2006).

9.3. Records from the Antarctic Margin

The Oligocene–Miocene boundary interval was first sampled in Antarctica at DSDP Site 270 in the Eastern Ross Sea (Fig. 9.1). Drilling recovered a succession of silty mudstones including glaciomarine sediment, which spans the Oligocene–Miocene boundary (Hayes and Frakes, 1975; Leckie and Webb, 1983). Although an abrupt lithological change with a potential hiatus at the Oligocene–Miocene boundary is noted by Leckie and Webb (1983), poor chronological resolution prevents unambiguous correlation with the Mi1 event and the earliest Miocene. The only exposed Oligocene–Miocene boundary strata reported from Antarctica crop out on King George Island (Fig. 9.1) and include the Destruction Bay Formation (Latest Oligocene) and Cape Melville Formation (earliest Miocene; Birkenmajer et al., 1985; Birkenmajer, 1987). In a recent summary of stratigraphy and facies of the succession, Troedson and Riding (2002) concluded that a significant glacial advance occurred at the boundary and that chronological control was good enough to suggest a correlation with the Mi1 event. The facies indicate a significant regional ice grounding event across Bransfield Strait and beyond the South Shetland Islands (Troedson and Riding, 2002). Unfortunately, drilling on the shelf and slope south of the South Shetland Islands (ODP Leg 178; Barker and Carmerlenghi, 2002) did not yield any more definitive records of the Oligocene–Miocene boundary.

Late Oligocene/early Miocene strata reported from the East Antarctic margin include Maud Rise (ODP Leg 113 sites 689 and 690; Barker et al., 1992), the Weddell Sea margin (ODP Leg 113 Site 693; Barker et al., 1992) and Kerguelen Plateau (ODP Leg 120 sites 747 and 748; Schlich and Wise, 1992; Fig. 9.1). The record at Maud Rise is relatively thin and comprises exclusively siliceous and carbonate ooze, although rare glacial drop stones are reported in strata from Site 689 (Barker et al., 1992). At the Weddell Sea margin Oligocene–Miocene sediments are also fine grained and include

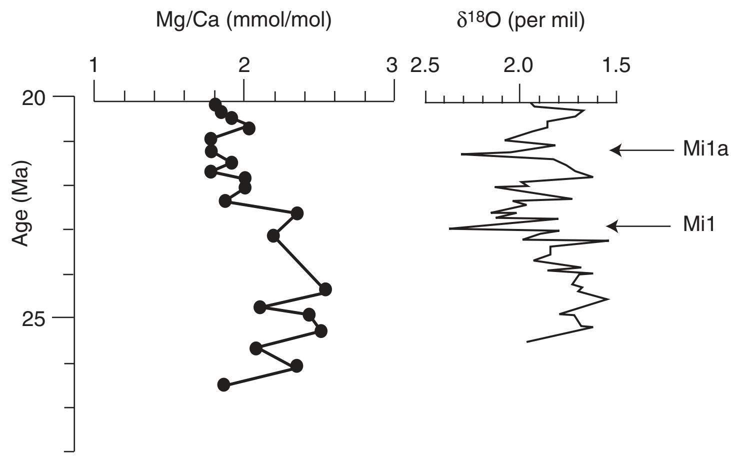

Figure 9.5: Oxygen Isotope and Mg/Ca Data from the Oligocene–Miocene Boundary Interval in ODP Site 747 from Kerguelen Plateau. Data from Billups and Schrag (2002) Adjusted to Time Scale of Billups et al. (2004).

diatom mud, clay and ooze (Barker ~~et al.,~~ 2002). Oligocene–Miocene boundary sediments at Kerguelen Plateau are also carbonate ooze (Schlich and Wise, 1992), however, foraminifera preservation and age resolution were good enough at ODP Site 747 to yield a benthic oxygen isotope stratigraphy across the boundary at that site (Wright and Miller, 1992; Billups and Schrag, 2002; Fig. 9.5)~~. However,~~ the amplitude of the Mi1 event was much reduced compared to equatorial values, with a δ^{18}O shift of only 0.3‰ across the Oligocene–Miocene boundary. A strontium isotope stratigraphy has also been assembled using planktic foraminifera from ODP Site 747 (Oslick et al., 1994). Oslick et al. (1994) reported significant increases in ^{87}Sr/^{86}Sr following the early Miocene Mi isotope events with a $\sim$1 myr lag. They suggested that this increase and similar subsequent stepwise increases in early–middle Miocene oceanic ^{87}Sr resulted from changes in the glacial state of East Antarctica.

Drilling in Prydz Bay (ODP Legs 119 and 188; Hambrey et al., 1991; Cooper and O'Brien, 2004) did not yield any Oligocene–Miocene age strata and Hambrey et al. (1991) concluded that this was due to erosion beneath an expanded middle-Miocene ice sheet.

9.3.1. McMurdo Sound, South Western Ross Sea

The Antarctic Oligocene–Miocene record is most complete in the Victoria Land Basin as recovered in the CIROS-1 and CRP-2A drill holes

(Barrett, 1989, Cape Roberts Science Team, 1999; Figs. 9.1 and 9.6). As with Prydz Bay, much of the Oligocene record of the Victoria Land Basin is marked by significant hiatuses (Wilson et al., 1998, 2000a, b), however, the latest Oligocene–early Miocene is preserved in both records (Naish et al., 2001; Wilson et al., 2002; Roberts et al., 2003). On the basis of radiometric, biostratigraphic and magnetostratigraphic data, Wilson et al. (2002) placed the Oligocene–Miocene boundary at 183.7 m in the CRP-2A core at the base of a normal polarity interval correlated with Polarity Subchron C6Cn.2n using Berggren et al.'s (1995) time scale. However, following the astronomical revision of the late Oligocene through early Miocene time scale (Billups et al., 2004; Gradstein et al, 2004; Pälike et al., 2004), Naish et al. (2008) placed the boundary at 130.27 m in an unconformity in the CRP-2A core and revised the age of strata underlying the unconformity to encompass Polarity Chron C7n. Roberts et al. (2003) placed the Oligocene–Miocene boundary at 274 m in the CIROS-1 core ~35 km south of CRP-2A. However, following the revision of Naish et al. (2008), the Oligocene–Miocene boundary in the CIROS-1 core more likely occurs in an unconformity at 92 m (Fig. 9.7). The boundary immediately overlies an unconformity at 248.71 m, which might represent as much as 1 myr following the age revision of Antarctic shelf diatom zones (Scherer et al., 2000) implied by Naish et al. (2008).

The strata recovered in both the CRP-2A and CIROS-1 drill holes are cyclic in nature and interpreted to represent periodic advance and retreat of ice across the Antarctic margin concomitant with sea-level fall and rise, respectively (Fielding et al., 1997; Naish et al., 2001). Each sequence is organised into a vertical succession, which begins with an erosion surface and is followed by a diamictite and sandstone, which gives way to sparsely fossiliferous bioturbated mudstone representing a cycle of glacial advance and retreat followed by open water conditions across the site of deposition (Naish et al., 2001) in concert with changes in relative sea-level (Dunbar et al., 2008; Fig. 9.6). Naish et al. (2008) estimated the glacioeustatic influence on relative water depth changes by deconvolving the tectonic, isostatic and palaeobathymetric components of water depth. These results are consistent with the δ^{18}O sea-level calibration of Pekar et al. (2002) from the New Jersey margin. Each cycle is interpreted to represent between 10 and 40 m of eustatic variation in the late Oligocene with perhaps 50 m of sea-level fall concomitant with an ice sheet some 20% larger than present coincident with the unconformity, which is correlated with the Oligocene–Miocene boundary by Naish et al. (2008). Grounding of ice across the site at the Oligocene–Miocene boundary is also confirmed by macro- and micro-structures indicative of glacio-tectonic deformation (Passchier, 2000;van der Meer, 2000).

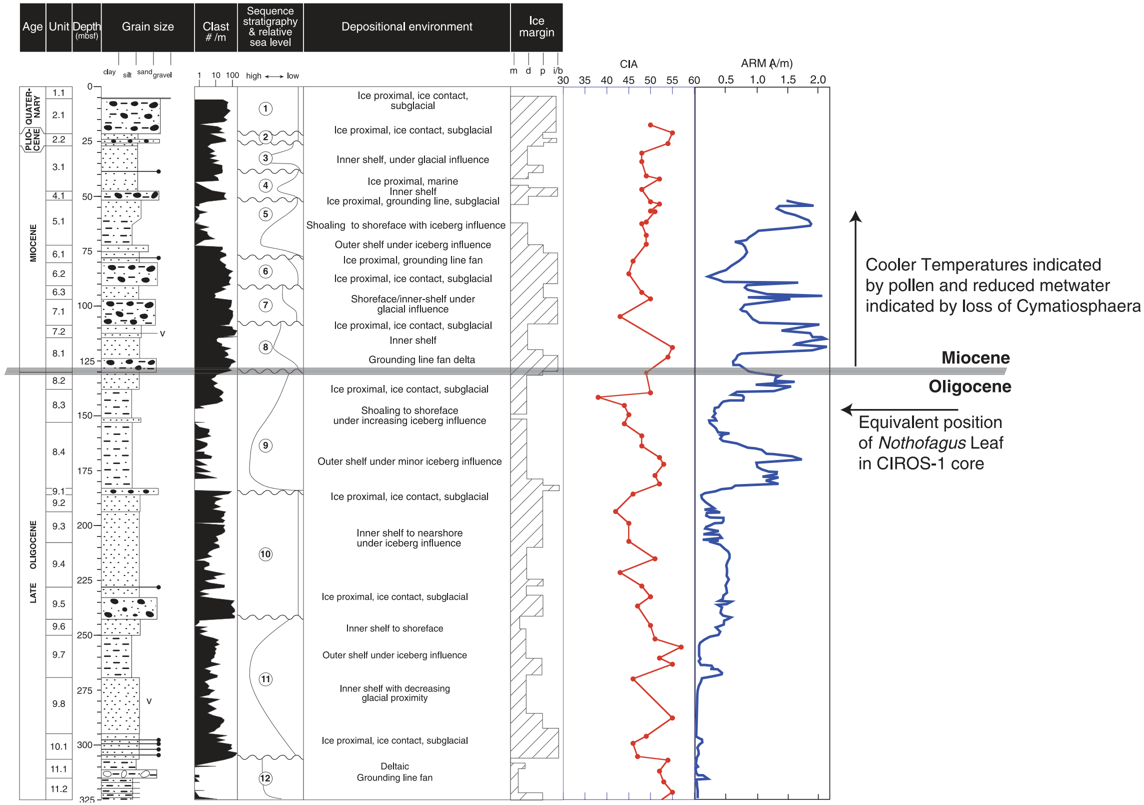

Figure 9.6: Environmental Proxy Data for the Upper Part of the CRP-2A Drill Core. Grain Size and Clast Data, and Sequence Stratigraphic, Palaeobathymetric, Depositional Environment and Ice Margin Interpretations are from Cape Roberts Science Team (1999). CIA, Chemical Index of Alteration (Data from Passchier and Krissek, 2008). ARM, Anhysteretic Remanent Magnetism (Data from Verosub et al., 2000). Temperature and Meltwater Indicators are Discussed in Barrett (2008). *Nothofagus* Leaf in CIROS-1 was Identified at 215.5 mbsf Immediately Underlying the Oligocene–Miocene Boundary as Identified from the Revised Age Model Presented in this Paper (Fig. 9.7).

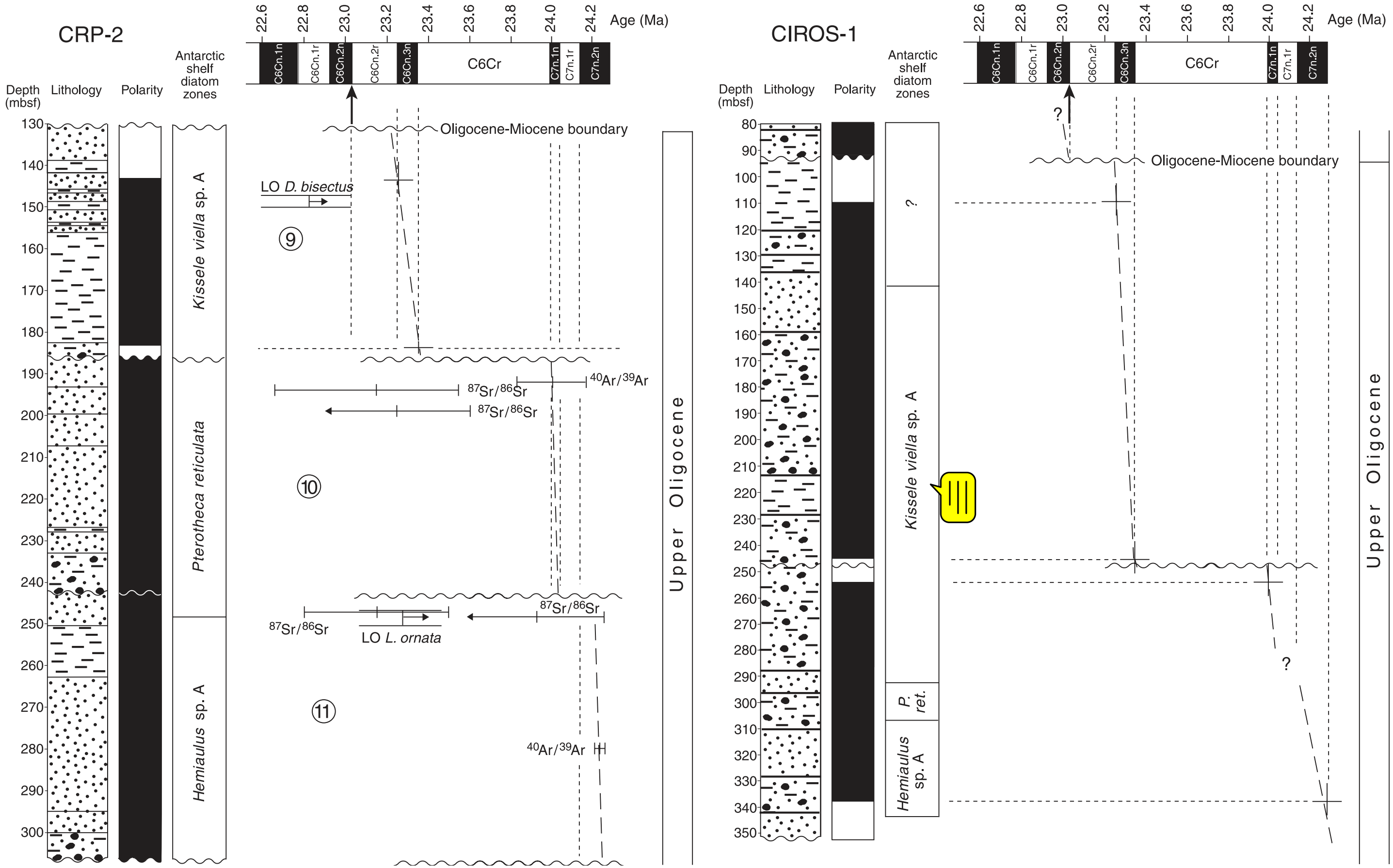

Figure 9.7: Revised Age Models for the CRP-2A and CIROS-1 Drill Cores from McMurdo Sound Using the Astronomical Time Scale of Billups et al. (2004) Following the Arguments in Naish et al. (2008). Magnetic Polarity and Diatom Biostratigraphy Data are from Wilson et al. (2000a, b, 2002) and Roberts et al. (2003). CIROS-1 Age Model is Constrained by Diatom Biostratigraphic Zones with Revised Ages from Naish et al. (2008).

The Oligocene–Miocene boundary mark~~s~~ some significant changes in physical properties in the CRP-2A core. Late Oligocene sedimentary cycles underlying the boundary are 55–60 m thick and relatively complete, whereas early Miocene sedimentary cycles are 10–20 m thick and truncated (Cape Roberts Science Team, 1999; Fig. 9.6). The clay mineralogy of the strata across the Oligocene–Miocene boundary in CRP-2A records stable physical weathering conditions (Ehrmann et al., 2005). Major element ratios derived from mudrock geochemistry for the same strata show significant shifts in the chemical index of alteration (CIA) across the Oligocene–Miocene boundary, which indicate periods of increased physical weathering and mechanical erosion associated with glacial advance (Passchier and Krissek, 2008). The CIA data reported by Passchier and Krissek (2008; Fig. 9.6) was corrected for the presence of primary volcanic detritus in order to reflect the palaeoclimatic record within the mudrock geochemistry. Short-lived glacial events at $\sim$23, $\sim$21 and $\sim$19 Ma indicated by the CIA data are correlated with the Mi events by Passchier and Krissek (2008) and interpreted to represent significant climatic and ice-sheet events in East Antarctica. Environmental magnetic properties also show a marked change across the Oligocene–Miocene boundary at 130.27 m in the CRP-2A core (Verosub et al., 2000). An earlier change in magnetic properties at 270 m (late Oligocene) is attributed to inception of the McMurdo Volcanic Province (Verosub et al., 2000).

Despite these changes in physical properties across the Oligocene–Miocene boundary in the CRP-2A core, palynological data although sparse due to low concentrations of organic matter, indicate a partially open landscape dominated by small *Nothofagus* (Southern Beech) stands or sparse tundra vegetation persisting through the late Oligocene–early Miocene (Askin and Raine, 2000). Oligocene–Miocene strata in the CIROS-1 core also contain similar amounts of pollen (Mildenhall, 1989) and a *Nothofagus* leaf fossil was preserved in latest Oligocene strata of the CIROS-1 Core. Mean summer temperature records derived from the (K+Na)/Al ratios of the CRP cores (Passchier and Krissek, 2008)~~, which~~ indicate~~s~~ relatively constant mean summer temperatures of $\sim$10°C in the latest Oligocene dropping to $\sim$6°C in the early Miocene. Marine palynomorphs, however, suggest a more significant change following the Oligocene–Miocene boundary with a significant reduction in the occurrence of prasinophyte algae, which is taken to indicate a reduction in offshore meltwater influence and hence cooler climates (Barrett, 2007).

9.4. Possible Drivers of Change Across the Oligocene–Miocene Boundary

9.4.1. Atmospheric Carbon Dioxide

While there has been no specific simulation of the influence of atmospheric CO_2 on ice-sheet growth at the Oligocene–Miocene boundary, DeConto and Pollard (2003a) demonstrated the potential link through simulations across the Eocene–Oligocene boundary. Their modelling demonstrated that ice-sheet inception occurred below a threshold of $3\times$ pre-industrial atmospheric CO_2 levels. Model results also demonstrate a strong response of ice volume to orbital forcing as atmospheric CO_2 approaches the glaciation threshold, and decreasing orbital variability of an established ice sheet as CO_2 approaches pre-industrial levels. Despite a predicted rewarming to pre-Oligocene–Miocene boundary levels in the ~~late Oligocene~~ and a major Antarctic glaciation in the middle Miocene (Zachos et al., 2001a), this was not matched by a parallel ~~increase~~ in levels of atmospheric CO_2 as determined by geochemical proxies (Pagani et al., 1999, 2005; Pearson and Palmer, 2000; Fig. 9.8). This apparent decoupling led Pagani et al. (1999, 2005) to conclude that, despite a significant decrease in atmospheric CO_2 from $\sim$500 ppmv in the late Oligocene to new near-modern values at the Oligocene–Miocene boundary, changing atmospheric CO_2 levels may have been secondary in driving Miocene Antarctic climatic and ice-sheet evolution.

9.4.2. Ocean Circulation/Tectonic Isolation

The progressive opening of oceanic gateways (Fig. 9.1) and progressive tectonic isolation during the Cenozoic stages of Gondwana breakup have been indicated as critical threshold events in the climatic deterioration and inception of ice sheets since the first deep-sea oxygen isotopic records were recovered (Shackleton and Kennett, 1975; Kennett, 1977). Testing of this hypothesis has proven particularly difficult due to uncertainties in the timing of gateway opening and inception of deep-water circulation. While estimates for the timing of the opening and deepening of the Tasmanian Gateway between Australia and Antarctica are reasonably well constrained to the Eocene–Oligocene boundary (Stickley et al., 2004), estimates for the opening of a deep Drake Passage range from the middle Eocene ($\sim$50 Ma) to the late

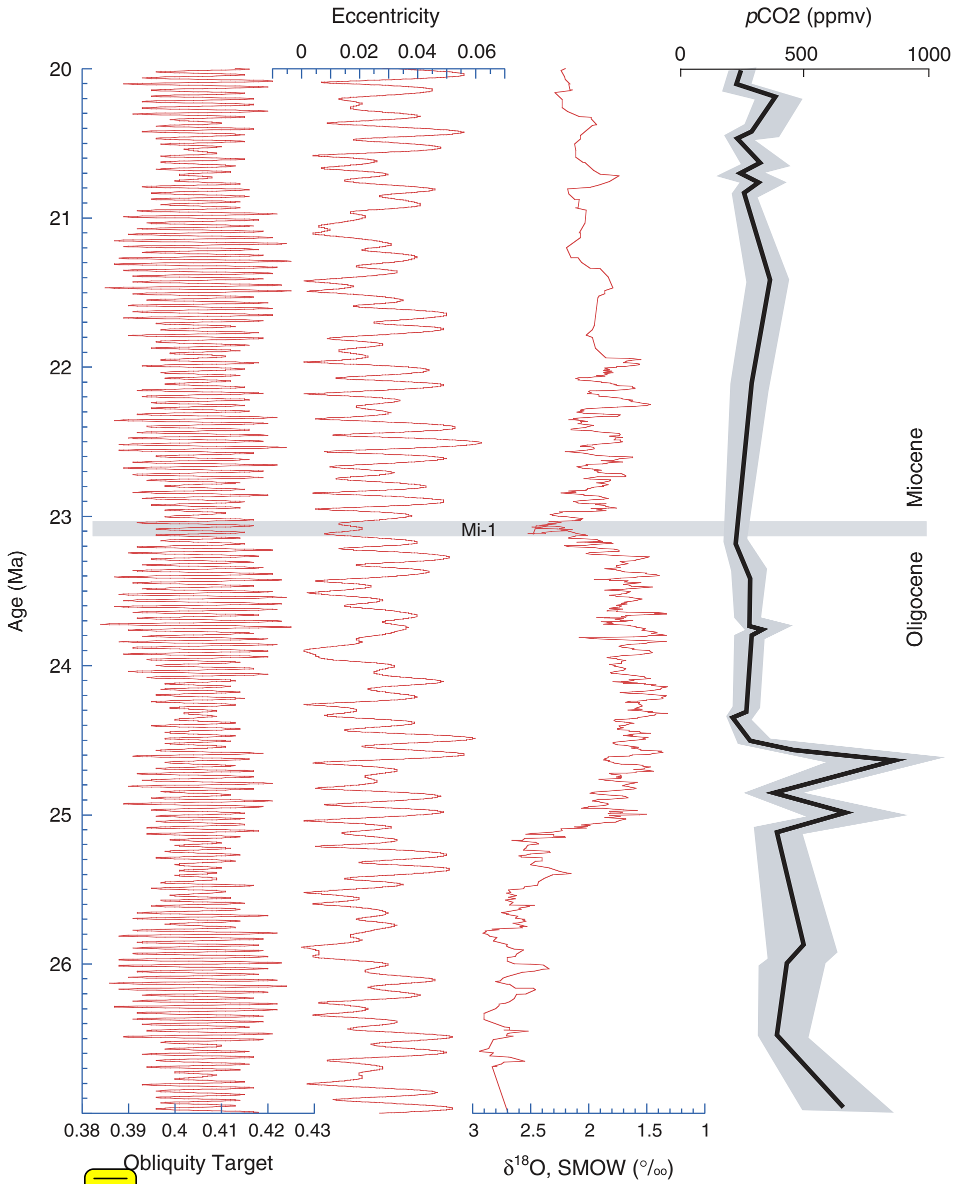

Figure 9.8: Potential Controls on Climate and Ice Sheet Behaviour across the Oligocene–Miocene Boundary and Origin of the Mi-1 $\delta^{18}O$ Event. Obliquity and Eccentricity Orbital Target Data from Laskar et al. (2003). $\delta^{18}O$ Data are from Zachos et al., (2001b) and pCO_2 Data are from Pagani et al. (2005). All Data are Plotted Against the Astronomical Time Scale Presented by Billups et al. (2004). See text for Discussion.

Miocene ($\sim$6 Ma) (Barker and Burrell, 1977; Barker et al., 2007; Livermore et al., 2007). However, some estimates suggest that deep-water circulation through the Drake Passage may well have been coincident with the Oligocene–Miocene boundary (Barker and Burrell, 1977, 1982). Pekar et al. (2006) and Pekar and Christie-Blick (2008) suggested, however, that Southern Ocean water masses were still relatively poorly mixed in the late Oligocene–early Miocene and that individual records used to compile the composite Cenozoic oxygen isotope curve by Zachos et al. (2001a) were drawn from water masses with different temperature and salinity histories. Hence, introducing an artifact from the splice of different isotopic records (Pekar and Christie-Blick, 2008) previously interpreted to represent significant oceanic warming in the latest Oligocene (Zachos et al., 2001a).

A number of numerical ocean modelling studies (Mikolajewicz et al., 1993; Nong et al., 2000; Toggweiler and Bjornsson, 2000;Sijp and England, 2004) have shown that the opening of a deep circum-Antarctic passage can cool the Southern Ocean by 1–3°C. While the amount of cooling in these studies is somewhat dependent on modelling details associated with the treatment of the atmosphere (Huber and Nof, 2006), the effects of this range of cooling on continental climate and ice-sheet mass balance have been shown to be small relative to the effects of the falling Cenozoic CO_2 concentrations (DeConto and Pollard, 2003a; Huber et al., 2004). For example, recent model simulations testing the importance of sea ice feedback on Antarctic ice sheets show that the continental interior is relatively insensitive to changes in Southern Ocean sea surface temperatures, and the effect of even large changes in ocean heat transport and sea ice is generally limited to the continental margins (DeConto et al., 2007). Conversely, the expansion of the EAIS, as presumed to have occurred at Mi1, has a dramatic effect on simulated Southern Ocean sea surface temperatures and sea ice distributions via the ice sheet's influence on regional temperatures and low-level winds (DeConto et al., 2007; Fig. 9.9). As these simulations clearly show, a growing Mi1 ice sheet would have cooled Southern Ocean sea surface temperatures by several degrees, pushing the 0 °C isotherm equatorwards and increasing the area, thickness, and fractional cover of seasonal and perennial sea ice (DeConto et al., 2007). Furthermore, as the katabatic wind field increased in intensity, the enhanced polar easterlies and westerlies would have increased ocean frontal divergence and upwelling, with possible implications for the marine carbon cycle and CO_2 drawdown (DeConto et al., 2007). Such mechanisms have been implicated as important contributors to the dynamics of Quaternary glacial cycles (Stephens and Keeling, 2000; Archer et al., 2003), but they have yet to be considered in a Miocene context.

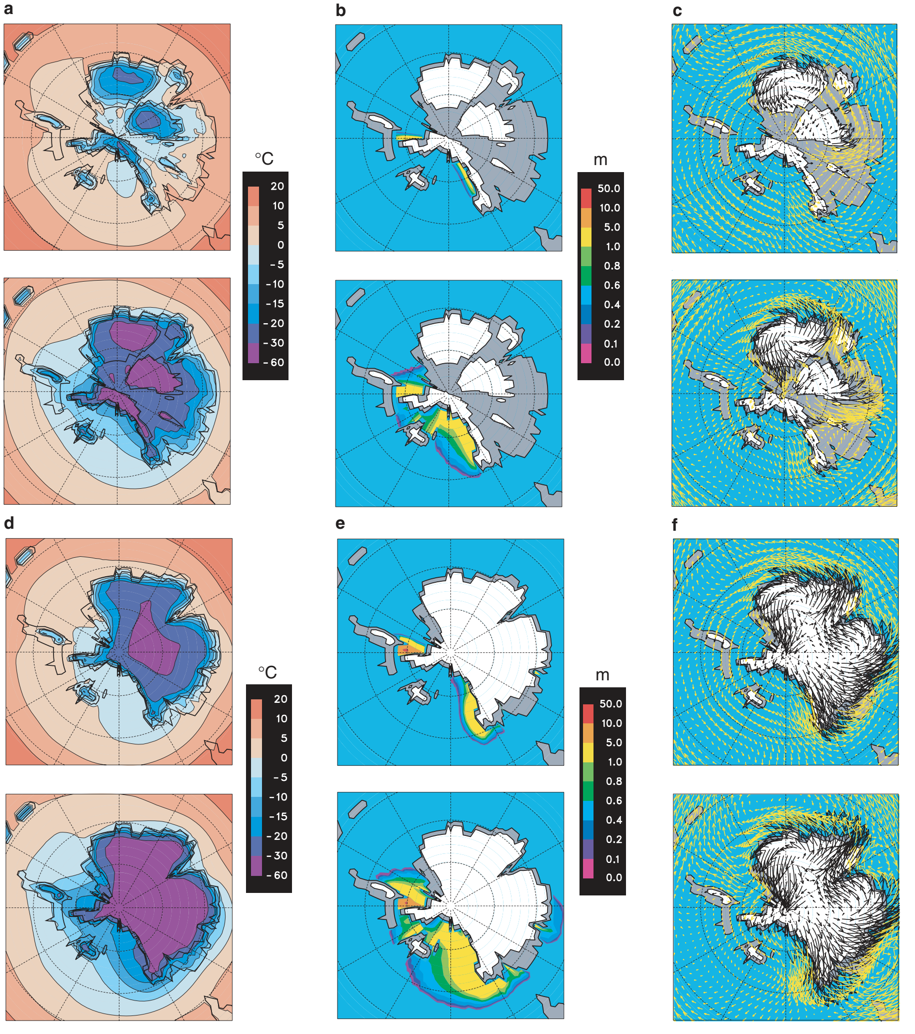

Figure 9.9: South Polar Seasonal Temperatures, Sea Ice and Winds in Response to a Growing Ice Sheet as Simulated by a GCM (DeConto et al., 2007). (a–c) Climatic Conditions in a Pre-Mi1 World with Isolated Ice Caps in the Continental Interior (Top). (d–f) Climatic Conditions with a Fully Glaciated East Antarctica as Presumed to have Existed at the Time of Mi1 (Bottom). With the Exception of Ice Sheet Geometry, Boundary Conditions are Identical in Both Simulations Including the Same Late Palaeogene Palaeogeography, 2 × Pre-Industrial CO_2 (560 ppmv), and a Relatively Cold Austral Summer Orbit Conducive to Antarctic Ice Sheet Growth. Ice Sheet Geometries are Taken from Prior GCM-Ice Sheet Simulations of Antarctic Glaciation (DeConto and Pollard, 2003a). Austral Summer (December, January, February) and Winter (June, July, August) Seasonal Climatologies are Shown on the Top and Bottom of (a–f), Respectively. (a, d) Seasonal Surface (2 m) Air Temperature, (b, e) Seasonal Sea Ice Extent and Thickness in Metres and (c, f) Lowest Level (Sigma Level = 0.189) GCM Winds with Vector Scale Length Equivalent to 2°C per $m-1$ of Wind Velocity.

While discussions of tectonic influences on Antarctic climate evolution usually focus on Southern Ocean gateways, Miocene ice sheets could have also been sensitive to changes in tropical climate associated with Tethyan tectonics. As the Southern Ocean gateways were widening and deepening, the eastern Tethys was closing. Ocean modelling studies have shown that the progressive closure of the Tethys affected the location of deep-water formation and the thermohaline component of the meridional overturning circulation, ocean heat transport, and both tropical and high-latitude sea surface temperatures (Hotinski and Toggweiler, 2003; von der Heydt and Dijkstra, 2006). While the Antarctic interior appears to ~~be~~ relatively insensitive to changes in the Southern Ocean, the modern Antarctic interior receives much of its moisture from the low mid-latitudes and significant changes in the tropics and associated teleconnections to polar latitudes could be important. Considering the timing of Tethyan closure relative to Antarctic ice sheet expansion in the Miocene, the sensitivity of ice-sheet evolution to low-latitude versus circum-Antarctic sea surface temperatures should be tested in future modelling studies.

The perspective provided by numerical climate modelling suggests falling greenhouse gas concentrations around the time of the Oligocene–Miocene boundary (Pagani et al., 2005) had a greater impact on Antarctic climate than the direct, physical effects of ocean gateways. However, the indirect effects of the gateways, including their influence on the marine carbon cycle and atmospheric CO_2 should also be considered. These indirect effects may be found to be more important to Cenozoic climate events like Mi1 than the direct influence of the gateways on ocean circulation and heat transport (Mikolajewicz et al., 1993; DeConto and Pollard, 2003a; Huber et al., 2004).

9.4.3. Orbital Parameters

If varying orbital parameters ~~are~~ to result in ice growth on the Antarctic craton, this ~~is~~ most likely to occur when summer insolation is minimised either by the seasonal timing of aphelion (precession) during periods of relatively high eccentricity, periods of low eccentricity minimis~~ing~~ the effects of precession, even when perihelion occurs during austral summer, or reduced amplitude in obliquity. All of these orbital configurations result in relatively cool summers and hence reduce the potential for summer melt. Zachos et al. (2001b) measured oxygen isotope ratios from Ocean Drilling Programme Equatorial Atlantic (Ceara Rise) Site 929 for a $\sim$5 myr interval spanning the Oligocene–Miocene boundary. The Mi1 event was found to correlate with both a minima in the low frequency (400 ky) eccentricity cycle and a prolonged minima in the amplitude of obliquity, a co-occurrence with a reoccurrence interval of 2.4 myr or longer (Pälike et al., 2004, 2006). This sustained period of unusually low seasonality (cold summers) was, however, relatively transient with warmer summers returning within a few hundred thousand years from the coincidence of increased eccentricity and high-amplitude variability in obliquity which would have resulted in warmer polar summers and increased summer melt (Zachos et al., 2001b).

9.4.4. Ice-Sheet Hysteresis

Coupled climate–ice sheet models have been reasonably successful in simulating sudden Cenozoic glaciation events such as Oi1 and Mi1 (DeConto and Pollard, 2003a, b). For example, beginning with an ice-free continent and assuming gradually declining greenhouse gas concentrations and accounting for orbital forcing, DeConto and Pollard (2003a) simulated the sudden stepwise glaciation of East Antarctica within a 200-ky interval. The simulated ice sheet was comparable in volume to the modern EAIS, but significantly smaller than the volume of Mi1 ice reconstructed from the proxy isotope and sea-level records discussed above. Subsequent modelling work, including a representation of ice shelves not included in their earlier simulations (Pollard and DeConto, 2007), have shown that an Antarctic ice sheet ~20–25% bigger than today would require a glaciated West Antarctica, and ice grounding lines extending close to the continental shelf break around much of the margin. An Mi1 ice sheet of this size would be similar in geometry to the ice sheet that existed at the Last Glacial Maximum (Huybrechts, 2002). However, the presumably warmer ocean at Mi1 would ~~be~~ unconducive to the seaward migration of grounding lines, so this scenario maybe difficult to reconcile from a modelling perspective. Furthermore, the cold south polar conditions implied by such an ice sheet impl~~y~~ global temperatures low enough to allow significant glaciation in the Northern Hemisphere, especially during orbital periods produc~~ing~~ cold boreal summers. While Greenland may have contained some glacial ice as early as the Eocene (Eldrett et al., 2007), the Oligocene–Miocene boundary is ~20 myr before the onset of the fist significant Northern Hemisphere glacial cycles. Clearly, some important problems remain in terms of reconciling the magnitude of the Mi1 event.

While the rapid growth of Antarctic ice at Mi1 can be explained through a combination of decreasing greenhouse gas concentrations and orbital forcing (with other possible influences from mountain uplift and/or ocean circulation), the ephemeral nature of the event and subsequent variability of ice volume are also problematic from a modelling perspective (Pollard and DeConto, 2005). As shown in both simple and sophisticated numerical ice-sheet models, (Weertman, 1961; Huybrechts, 1994; Pollard and DeConto, 2005), the high albedo and elevation of large polar-centred ice sheets produce considerable hysteresis. In a scenario of cooling climate, a polar ice sheet can grow suddenly, once the snow line intersects sufficient land area in mountains and high plateau. The non-linear jump in ice volume is facilitated by height–mass balance and albedo feedbacks, as the ice sheet spreads horizontally (albedo feedback) and more of the parabolic ice surface rises above the snow line and out of the ablation zones around its margins (height–mass balance feedback) (Abe-Ouchi and Blatter, 1993; DeConto and Pollard, 2003a). The high elevation and albedo of the ice sheet inhibit the ice sheet from disappearing during subsequent warming interval, unless temperatures (snow lines) rise far above their initial values (elevation) at the time of glacial onset (Huybrechts, 1994). Pollard and DeConto (2005) studied this hysteresis effect in a coupled GCM–ice sheet model and in a simple flowline model with parameterised mass-balance forcing. They concluded that the hysteresis effect is strong enough to preclude orbital forcing from driving the range of Cenozoic ice-volume variability seen in the oxygen isotope and sea-level records described above, unless the orbital forcing is accompanied by significant changes in greenhouse concentrations. During favourable (cold austral summer) orbital periods, the atmospheric CO_2-glaciation threshold for Antarctica is $\sim 2\times$ pre-industrial levels, while CO_2 must approach $\sim 4\times$ pre-industrial levels during a warm austral summer orbital period to trigger the collapse of the interior EAIS. If the sensitivity of the models to orbital and greenhouse gas forcing is reasonable, the short duration of the peak Mi1 event would require a significant perturbation to the carbon cycle, producing significant global warming soon after the peak glacial interval. Greenhouse gas variability of this magnitude is not ~~clearly~~ evident in existing proxy reconstructions of early Miocene CO_2 (e.g. Pagani et al., 2005), however, higher-resolution records will be required to resolve this type of CO_2 variability across key climatic events like the Oligocene–Miocene boundary.

9.5. Summary and Conclusions

Strata recovered from the Antarctic margin indicate a significant glacial advance at the Oligocene–Miocene boundary reaching the south Shetland Islands on the Antarctic Peninsula (Troedson and Riding, 2002) and grounding in Prydz Bay and the South Western Ross Sea as indicated by hiatuses in drill cores (Hambrey et al., 1991; Roberts et al., 2003; Naish et al., 2008). Ice rafted as far north as Maud Rise (Barker et al., 1988a, b) and the central Ross Sea (Leckie and Webb, 1983) but did not appear to reach the Kerguelen Plateau (Schlich and Wise, 1992). Pre-Oligocene–Miocene boundary strata indicate a late Oligocene Antarctic ice sheet (Cape Roberts Science Team, 1999), which expanded to an ice volume of the order of 20% greater than the present ice sheet at the Oligocene–Miocene boundary (Naish et al., 2008). The glacial expansion, however, although significant in extent and volume, must have been relatively transient and neither cold nor extensive enough to extinguish *Nothofagus* tundra vegetation (Askin and Raine, 2000; Roberts et al., 2003), which persisted across the boundary despite a slight drop in temperature (Passchier and Krissek, 2008). Marine palynomorphs, however, indicate that coastal temperatures did not return to the warmth of the late Oligocene with a much reduced freshwater melt input to coastal regions (Barrett, 2007).

Data from the Antarctic continent are entirely consistent with the short-lived (200 ky) ice-volume increase from 40% of present Antarctic ice volume to 25% greater than present Antarctic ice volume across the Oligocene–Miocene boundary with concomitant oceanic deep-water cooling implied by the Mi1 isotopic excursion recognised in equatorial and Southern Hemisphere deep-sea sedimentary records (Paul et al., 2000). Ice-volume estimates are confirmed by the backstripped stratigraphic records from the New Jersey Margin (Kominz and Pekar, 2001; Pekar et al., 2002), however, accommodating this much ice on Antarctica when global temperatures were presumably warmer than today may prove difficult from a modelling perspective. Warm summer mean temperatures were re-established soon after the Oligocene–Miocene boundary, although a few degrees cooler than pre-Miocene summer mean temperatures. The duration and transience of the Mi1 glacial expansion and swift recovery in Antarctica likely resulted from the limited polar summer warmth from coincidence of low eccentricity and low-amplitude variability in obliquity of the earth's orbit at the Oligocene–Miocene boundary (Zachos et al., 2001b). This was followed by warmer polar summers and increased melt from increased eccentricity and high-amplitude variability in obliquity in the early Miocene, allowing the recovery of vegetation on the Antarctic craton. Atmospheric CO_2 concentrations remained below the $2\times$ pre-industrial threshold, which promoted sensitivity of the climate system to orbital forcing during cold Austral summers. While climate and ice-sheet modelling supports the fundamental role of greenhouse gas forcing punctuated by orbital forcing as a likely cause of events like Mi1 (DeConto and Pollard, 2003a; Huber et al., 2004; DeConto et al., 2007;

Pälike et al., 2007), the models underestimate the range of orbitally paced ice-sheet variability recognised in early Miocene isotope and sea-level records unless accompanied by significant fluctuations in greenhouse gas concentrations (Pollard and DeConto, 2005). While, tectonic influence may have been secondary, they may well have contributed to oceanic cooling recorded at the Cape Roberts Project site in the South Western Ross Sea (Barrett, 2007).

Uncited Reference

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