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Resolving a late Oligocene conundrum: Deep-sea warming and Antarctic glaciation

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Abstract

Changes in ice volume and resulting changes in sea level were determined for the late Oligocene (26–23 Ma, Astronomical Timescale, ATS) by applying δ^{18}O-to-sea-level calibrations to deep-sea δ^{18}O records from ODP Sites 689, 690, 929, 1090, and 1218. Our results show that maximum global ice volume occurred during two late Oligocene δ^{18}O events, Oi2c (24.4 Ma) and Mi1 (23.0 Ma) (inferred glacioeustatic lowering), with volumes up to ~25% greater than the present-day East Antarctic Ice Sheet (EAIS). Ice volume during glacial minima was on the order of about 50% of the present-day EAIS. This is supported by late Oligocene stratigraphic records from Antarctica that contain evidence of cold climates and repeated episodes of glaciation at sea level and grounding lines of glacial ice on the Antarctic continental shelf in the Ross Sea and Prydz Bay. In contrast, composite deep-sea δ^{18}O records show a significant decrease (≥ 1‰) between 26.7 and 23.5 Ma, which have long been interpreted as bottom-water warming combined with deglaciation of Antarctica. However, a close examination of individual δ^{18}O records indicates a clear divergence after 26.8 Ma between records from Southern Ocean locations (i.e., Ocean Drilling Program Sites 689, 690, 744) and those of other ocean basins. High δ^{18}O values (2.9‰–3.3‰) from these Southern Ocean δ^{18}O records are consistent with an ice sheet on the East Antarctic continent equivalent to present-day values and cold bottom-water temperatures (≤ 2.0 °C). These differences suggest a reduction in deep-water produced near the Antarctic continent (i.e., proto-Antarctic Bottom Water, proto-AABW), which were quickly entrained and mixed with warmer (and presumably more saline) bottom-water originating from lower latitudes. Expansion of a warmer deep-water mass and the weakening of the proto-AABW may explain the large intra-basinal isotopic gradients that developed among late Oligocene benthic δ^{18}O records. These conclusions are also supported by ocean modeling suggesting a reduction of deep-water formed in the Southern Ocean, strengthening of deep-water from the northern hemisphere, and decreasing temperatures in high southern latitudes occurred as the Drake Passage opened to deep-water. Low δ^{18}O values reported from deep-sea locations other than the Southern Ocean are shown to bias estimates of Antarctic ice volume, calling for a re-evaluation of the notion that Antarctic ice volume was significantly reduced during the late Oligocene.

Keywords: Oligocene; Oxygen isotopes; Sea-level; Ice volume; Antarctica

1. Introduction

For decades, palaeoceanographers observed a significant decrease in δ^{18}O values in late Oligocene com-

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posite deep-sea records (e.g., Miller et al., 1987; Zachos et al., 2001), which was generally attributed to deep-sea warming, combined with a significant decrease in Antarctic ice volume (e.g., Zachos et al., 2001). This interpretation is supported by the assumption that bottom-water emanated mainly from a Southern Ocean source (i.e., Southern Component Water, presumably originating around the Antarctic margin) during the Oligocene (e.g., Wright and Miller, 1993); thus, serving as a proxy for palaeoenvironmental conditions on the Antarctic continent. Isotopic records from recently cored ODP sites (Sites 1090 and 1218) contain higher values, but are still near the threshold for requiring the presence of an ice sheet (2.45‰; Miller et al., 1987).

In contrast, results from recent stratigraphic drilling of the Antarctic continental margin (e.g., CIROS-1, Cape Roberts Project) indicate a gradual and steady Antarctic cooling during the Oligocene, culminating in near tundra-like conditions by the early Miocene (e.g., Raine, 1998; Raine and Askin, 2001; Thorn, 2001; Roberts et al., 2003; Prebble et al., this issue). Palaeoenvironmental evidence from these terrestrial palynological and phytolith data, as well as the sedimentary record from glacigenic sediments recovered in upper Oligocene strata from the western Ross Sea indicates that Antarctica was sufficiently cold to support the existence of ice sheets at sea level (e.g., Barrett, 1989; Cape Roberts Science Team, 1998, 1999, 2000; Naish et al., 2001). Identification of ice grounding lines near the shelf edge near Prydz Bay as early as the early Oligocene (Cooper et al., 1991; Bartek et al., 1997) also suggests the presence of large continental ice sheets on East Antarctica at this time. Additionally, eustatic estimates from sequence stratigraphic records (Kominz and Pekar, 2001) indicate repeated sea-level lowerings during the late Oligocene (25.1–23.0 Ma) consistent with a heavily glaciated East Antarctic continent (EAC) (Pekar et al., 2002).

This paper addresses the paradox of low $\delta^{18}O$ values in deep-sea records coeval with proximal Antarctic records suggesting decreasing and persistent cold temperatures and large-scale ice sheets on East Antarctica. Here we show that while ice volume may have fluctuated on orbital timescales, Antarctica could have remained mostly glaciated (equivalent to ~50% to 125% of present-day EAIS) throughout the late Oligocene. We also suggest that the apparent late Oligocene warming interpreted from deep-sea $\delta^{18}O$ records could have been caused by an expansion of warmer deep-water into most of the world's ocean basins, with colder deep-water becoming entrained with and mixed into this warmer deep-water mass.

2. Methods, definitions, and sites used in this study

Oxygen isotope records from DSDP and ODP Sites 522 (Miller et al., 1988), 529 (Miller et al., 1991), 558 (Miller and Fairbanks, 1983), 563 (Miller and Thomas, 1985), 689 (Kennett and Stott, 1990), 690 (Kennett and Stott, 1990), 744 (Zachos et al., 2001), 747 (Wright et al., 1992), 748 (Zachos et al., 1992), 754 (Zachos et al., 2001), 803 (Barrera et al., 1993), 929 (Zachos et al., 2001), 1090 (Billups et al., 2002), and 1218 (Lear et al., 2004) were used in this study (Fig. 1). Chronologies for these records were previously developed by integrating biostratigraphy and magnetostratigraphy, which have been converted to the new Astronomical Time Scale (ATS) of Laskar et al. (2005). These original age models vary in their uncertainties, which can affect the development of time slice isotopic transects. However,

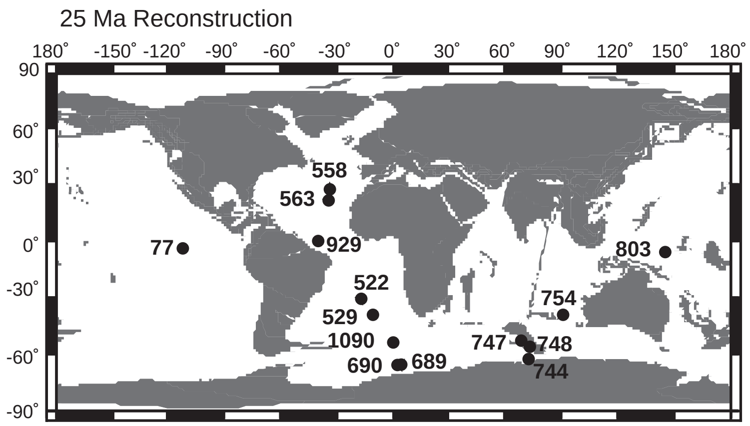

Fig. 1. Map showing plate tectonic reconstructions at circa 25 Ma (from Hay et al., 1999) with the palaeo-locations of DSDP and ODP sites used in this study.

each time slice includes a 400 ky interval across each δ^{18}O event, which should be a sufficient length of time to capture isotopic values representative of the event.

For Sites 689 and 690, which contain two important isotopic records for the interpretations in this study, Florindo and Roberts (2005) provided new chronostratigraphic interpretations using u-channel samples for the upper Oligocene. Although their new age models do not significantly alter late Oligocene ages for Site 689, ages near the top of the Oligocene section in Site 690 are approximately 0.3 m.y. younger than previous age models (e.g., Zachos et al., 2001). In Florindo and Roberts (2005), the top of Chron 8n (25.1 Ma, ATS) has been assigned to a change from normal to reversed polarity at 55.0 mbsf. Therefore, an interval of normal polarity occurring between 54.8 and 53.2 mbsf (Speiss, 1990) is assigned to Chron 7a (25.0–24.8 Ma) and a short interval of normal polarity that begins immediately below an unconformity at ~50 mbsf is assigned to Chron 7n (24.5–24.1 Ma). The last isotope measurement at Site 690 is obtained from a sample taken at 50.75 mbsf, which occurs within the lower portion of Chron 7n and is therefore assigned an age of ~24.5 Ma.

Deep-sea isotopic values are typically obtained for a single species, such as *Cibicidoides* spp. However, *Cibicidoides* spp as do most other benthic foraminifers precipitate their tests out of equilibrium with calcite (δ^{18}O values in the tests of *Cibicidoides* spp. are offset with respect to calcite by + 0.64‰). All δ^{18}O records have been adjusted to represent the isotopic value of calcite for samples reported either by previous authors (i.e., Zachos et al., 2001) or in this study (i.e., Site 1090).

The term apparent sea level (ASL) is used here and is defined as eustasy plus the water-loading effects on the crust (eustasy * ~1.48; Pekar et al., 2002). The maximum size of a fully glaciated East Antarctic continent during the Oligocene is estimated to be equivalent to ≤ 80 m ASL. This is based on δ^{18}O to ASL calibrations (Pekar et al., 2002), coupled δ^{18}O and Mg/Ca ratio records by Lear et al. (2000), and two-dimensional flexural backstripping and stratigraphic studies (Kominz and Pekar, 2001). This ice volume estimate is somewhat greater than recent GCM-ice sheet simulations of the Oi1 event (DeConto and Pollard, 2003a), which ignored significant ice on West Antarctica and the seaward expansion of grounding lines beyond the model shorelines, and is closer to simulations of maximum Antarctic ice volume during Quaternary glacial periods (Ritz et al., 2001), when the total area of grounded ice on East and West Antarctica was 15%–25% greater than today (Denton and Hughes, 2002; Huybrechts, 2002). Calibrations of δ^{18}O to ASL amplitudes use detrended ASL estimates from Pekar et al. (2002) and benthic foraminiferal δ^{18}O amplitudes at Oi- and Mi-events (Miller et al., 1991; Pekar and Miller, 1996) from ODP Sites 689, 690, 744, 929, and 1218 (Fig. 2; Pekar et al., 2002). Detrended ASL estimates were derived by integrating two-dimensional flexural backstripping (Kominz and Pekar, 2001) with two-dimensional palaeoslope modeling of foraminiferal biofacies and lithofacies (Pekar and Kominz, 2001). Lowest calibrations are from the Weddell Sea ODP Sites 690 and 689 (0.12‰/10 m ASL, $r^2=0.92$ and 0.13‰/10 m ASL, $r^2=0.72$, respectively), with higher values for Sites 1218 (0.16‰/10 m ASL, $r^2=.67$) and 744 (0.26‰/10 m ASL, $r^2=0.82$) (Fig. 2). The calibrations for Sites 929 and 1090 use a single δ^{18}O event (Mi1, 23.0 Ma). This results in a calibration ranging from ~0.2‰ to 0.5‰ (0.35‰ ± 0.15‰ mean calibration) and 0.18‰ to 0.46‰ (0.32‰ ± 0.14‰ mean calibration) per 10 m ASL for Sites 929 and 1090, respectively, based on an ASL estimate of 56 ± 25 m. Differences among these calibrations are attributable to greater variability in deep-sea temperatures between glacial maxima and minima at the million-year timescale. The temperature signal within the observed isotopic shifts ranges from 25% at Site 690 to ~75% at Site 929, which is estimated by subtracting the ice volume contribution (estimated to be 0.091‰/10 m ASL, DeConto and Pollard, 2003a) from δ^{18}O to sea-level calibrations.

For estimating Oligocene ice volume, δ^{18}O values of 3.0‰ or greater in deep-sea records are consistent with a fully glaciated EAIS and cold bottom-water temperatures (~2.0 °C). This is based on the following. The modern *Cibicidoides* spp. value in ~2.0 °C water is 2.7‰ (Shackleton and Kennett, 1975) or 3.34‰ adjusted for equilibrium. Of that value, the isotopic contribution from the present-day ice sheets is estimated to range from ~0.9‰ to 1.2‰ (e.g., Miller et al., 1991; Zachos et al., 2001). In this study, the isotopic contribution of the present-day ice sheets is estimated to be ~1.0‰, based on present-day ice volume estimates from Williams and Ferrigno (1999) (Table 1), using both grounded ice and ice below sea level. Of that 1.0‰ value, 0.13‰ is attributed to ice from Greenland and West Antarctica (Table 1). This value includes average isotopic values of the present-day West Antarctic and Greenland ice sheets of approximately − 30‰ and − 39‰, respectively (obtained by taking the average values of δ^{18}O records from ice cores for each area). During the late Oligocene, Greenland and West Antarctica may not have been glaciated, reducing

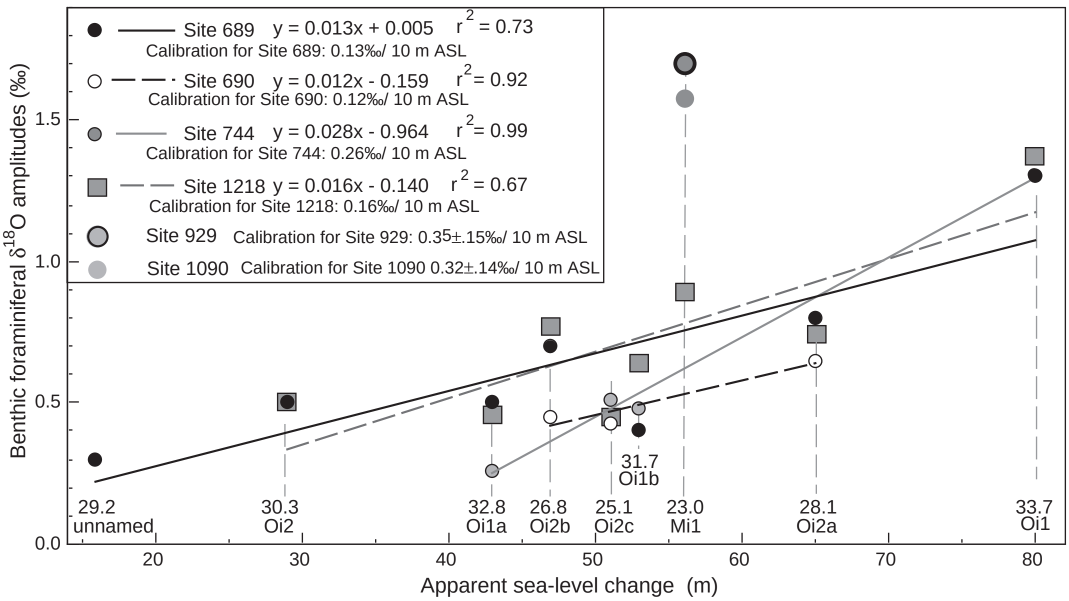

Fig. 2. Oxygen isotope amplitudes from Oi-events identified in ODP Sites 690, 689, 744, 929, 1090, and 1218 (Thomas et al., 1995; Zachos et al., 2001; Lear et al., 2004) are compared to detrended ASL amplitudes (Pekar et al., 2002). Oxygen isotope amplitudes are the difference between the maximum value of δ^{18}O event and preceding minimum (or average minimum) δ^{18}O value. ASL amplitudes (Pekar et al., 2002) are the difference between sea-level minimum and the preceding sea-level maximum. Oxygen isotope events and their ages are shown (Miller et al., 1991; Pekar and Miller, 1996). These correlations suggest that although benthic foraminiferal records are assumed to contain a significant bottom-water temperature signal, lowering of temperature results in a nearly linear increase in ice volume estimates. The δ^{18}O amplitude from Sites 929 and 1090 is calibrated to ASL with a single δ^{18}O event, which results in a larger uncertainty. It is also uncertain whether these calibrations would remain constant between high frequency (e.g., obliquity timescales) and million-year timescale (e.g., Oi- and Mi-events of Miller et al., 1991).

this average deep-sea δ^{18}O value by 0.13‰. Furthermore, a wet-based, polythermal East Antarctic ice sheet, such as thought to have existed during the Oligocene, likely had significantly higher δ^{18}O values in the ice sheet (e.g., ~− 35‰) than values today (i.e., − 45‰ to − 55‰). We estimate that this would have further reduced the isotopic value contribution of ice during the Oligocene by 0.25‰ (Table 1). This in turn would have resulted in an isotopic value of calcite of ~3.0‰ for the Oligocene, which is consistent with 2 °C bottom-water concomitant with a fully glaciated East Antarctic continent. Uncertainty in this value includes the possible effects of salinity. The range in salinity in the deep-sea today (e.g., North Atlantic Deep Water

Table 1
Ice volume and resulting isotopic change to mean global ocean if the ice sheet melted

Ice sheet	Ice volume (km^3)	Water volume (km^3)	Isotopic change (‰)		
			Present-day isotopic values	Isotopic value if all ice is − 35‰	Isotopic value if all ice is − 50‰
Greenland	2,600,000	2,340,000	− 0.07	− 0.06	− 0.09
East Antarcitc Ice sheet	26,039,200	23,435,280	− 0.85	− 0.60	− 0.85
West Antarcitc ice sheet	3,262,000	2,935,800	− 0.06	− 0.07	− 0.11
Other Antarctic ice	808,600	727,740	− 0.02	− 0.02	− 0.03
Other ice	180,000	162,000	0.00	0.00	− 0.01
Total	32,889,800	29,600,820	− 1.00	− 0.75	− 1.08

δ^{18}O value of calcite with present-day ice volume and a bottom-water temperature of 2.0 °C (Shackleton and Kennett, 1975).	3.34
Isotopic increase to oceans from Greenland and WIAS only	− 0.13
δ^{18}O value of deep-sea calcite with EAIS present and a bottom-water temperature of 2.0 °C.	3.21
Difference between present-day and Oligocene ice sheet isotopic contribution to ocean	− 0.25
δ^{18}O value of calcite an Oligocene EAIS and a bottom-water temperature of 2.0 °C.	2.96

Note: that the following data are used here: the surface area of the ocean=~362,000,000 km^2, the average ocean depth=~3.8 km, and ocean volume=~1,375,600,000 km^3. Ice volume and water volume estimates are from Williams and Ferrigno (1999).

[34.95‰] and Antarctic Bottom Water [34.65‰]) is about 0.3‰, resulting in uncertainty in δ^{18}O of approximately ± 0.1‰. A warmer water mass with higher salinity would have a higher δ^{18}O, resulting in an over-estimate in ice volume. In contrast, a colder deep-water mass with lower salinity, such as a proto-AABW, would have had a lower δ^{18}O, resulting in an under-estimate of the ice volume. Recent studies have suggested that West Antarctica and perhaps even northern hemisphere continents may have been glaciated during the Oligocene (e.g., Coxall et al., 2005). However, if a West Antarctic Ice Sheet (WAIS) existed during the Oligocene, its isotopic contribution would be minor (0.06‰), and the presence of northern hemisphere ice sheets during the Palaeogene has little evidence from the northern hemisphere fauna, flora, or lithological studies to support it (e.g., Wolfe, 1978; Wolfe and Poore, 1982; Axelrod and Raven, 1985; Tiffney, 1985). Even so, a fully developed Greenland ice sheet would increase the isotopic value of the global oceans by 0.07‰ and combined with a WAIS could be invoked to explain a small fraction of the larger ice volume estimates recently proposed for the Oligocene (equivalent to 80 m ASL, Kominz and Pekar, 2001; 90 m ASL Lear et al., 2004; 100 to 160 m ASL, Coxall et al., 2005).

Apparent sea-level estimates were calculated using the δ^{18}O to ASL calibrations for Sites 689, 690, 744, 929, and 1218 (Pekar et al., 2002; this study; Fig. 2). Isotopic values of ≥ 3‰, which occur at each late Oligocene isotopic event, are used to indicate an ice sheet equivalent in size to the present-day EAIS. The calibrations are further refined at each isotopic event by comparing isotopic offsets among the δ^{18}O records, which is ascribed to temperature variability between ocean basins. For example, at Mi1, Sites 929 and 1218 are ~0.4‰ lighter than Sites 1090, which is assumed to be due to cooler bottom-water bathing Site 1090.

Although previous calibrations indicate that temperature scales linearly with respect to ice volume, uncertainties in temperature variability still may exist. For example, bottom-water temperature changes could occur outside the variability suggested by the calibrations for a given site owing to long or short-term changes in deep-sea circulation patterns. Additionally, apparent sea-level/ice volume estimates from high δ^{18}O values (≥ 3‰) are considered more robust, because these values are consistent with a fully glaciated East Antarctic continent and cold bottom-water temperatures, while lower δ^{18}O values could have a wider range of possible ice volume and bottom-water temperatures. For example, the highest values at Site 1090 are consistent with an ice sheet up to 25% larger than the present-day EAIS and bottom-water temperatures near or slightly colder than water temperatures currently bathing Site 1090 (Billups et al., 2002). In fact, to invoke a smaller ice volume would require the unlikely scenario of colder bottom-water occurring during the late Oligocene, a time with a warmer climate and a polythermal ice sheet, in contrast to the extreme cold polar conditions that exist in Antarctica today.

3. Deep-sea δ^{18}O records from the late Oligocene

Deep-sea δ^{18}O records in most oceanic basins show a significant (> 1.0‰) decrease after the δ^{18}O event (Oi2b) at 26.8 Ma, reaching their lowest values of the Oligocene by ~24.5 Ma (Fig. 3). For example, low-resolution δ^{18}O records from Atlantic Sites 522, 529, 558, 563 (Miller et al., 1987, 1991) all indicate a decrease between 26.6 and 23.5 Ma, culminating with low δ^{18}O values between ~2.0‰ and 1.2‰ (Fig. 3). Low values in the high-resolution tropical Atlantic Ocean ODP Site 929 record extend from 25.2 Ma to immediately before the Mi1 event. A similar trend occurs in δ^{18}O record from tropical Pacific Ocean ODP Site 1218, with high δ^{18}O values (i.e., glacial maxima) decreasing by ~1‰ (Lear et al., 2004). These values are heavier than Site 929 on average by 0.5‰ to 0.8‰, during δ^{18}O maxima and minima, respectively. Other Pacific Ocean ODP Sites 77 and 803 show a decrease in δ^{18}O values of ~1‰ during the late Oligocene. These low δ^{18}O values have been used to suggest a warming event occurred during the late Oligocene coupled with a collapse of the Antarctic ice sheet (e.g., Zachos et al., 2001). These isotopic values are similar to middle Eocene values, a time usually considered to be mainly ice free (Zachos et al., 2001). Slightly higher values δ^{18}O (1.8‰ to 2.6‰) are recorded in South Atlantic Site 1090 (Billups et al., 2002) between 23.8 and 23.0 Ma, with the highest values just above the threshold requiring the presence of ice sheets (2.45‰; Miller et al., 1991).

Unlike the low δ^{18}O values observed in records in the Atlantic and Pacific Ocean basins, δ^{18}O values from Southern Ocean ODP Sites 689, 690, and 744 remain high (~3‰). In the case of Site 690, they increase to 3.1‰–3.3‰ in the upper portion of the record (25.2–24.5 Ma), with δ^{18}O records from Sites 744 and 689 also containing relatively high values (2.9‰–3.0‰) at the top of their records (24.9 and 25.4 Ma, respectively). In contrast, between ~26.0 and 24.5 Ma, an isotopic gradient of ~0.6‰ to 1.6‰ developed between Site 690 (2.9‰–3.3‰) and North Atlantic Sites 529, 558, and

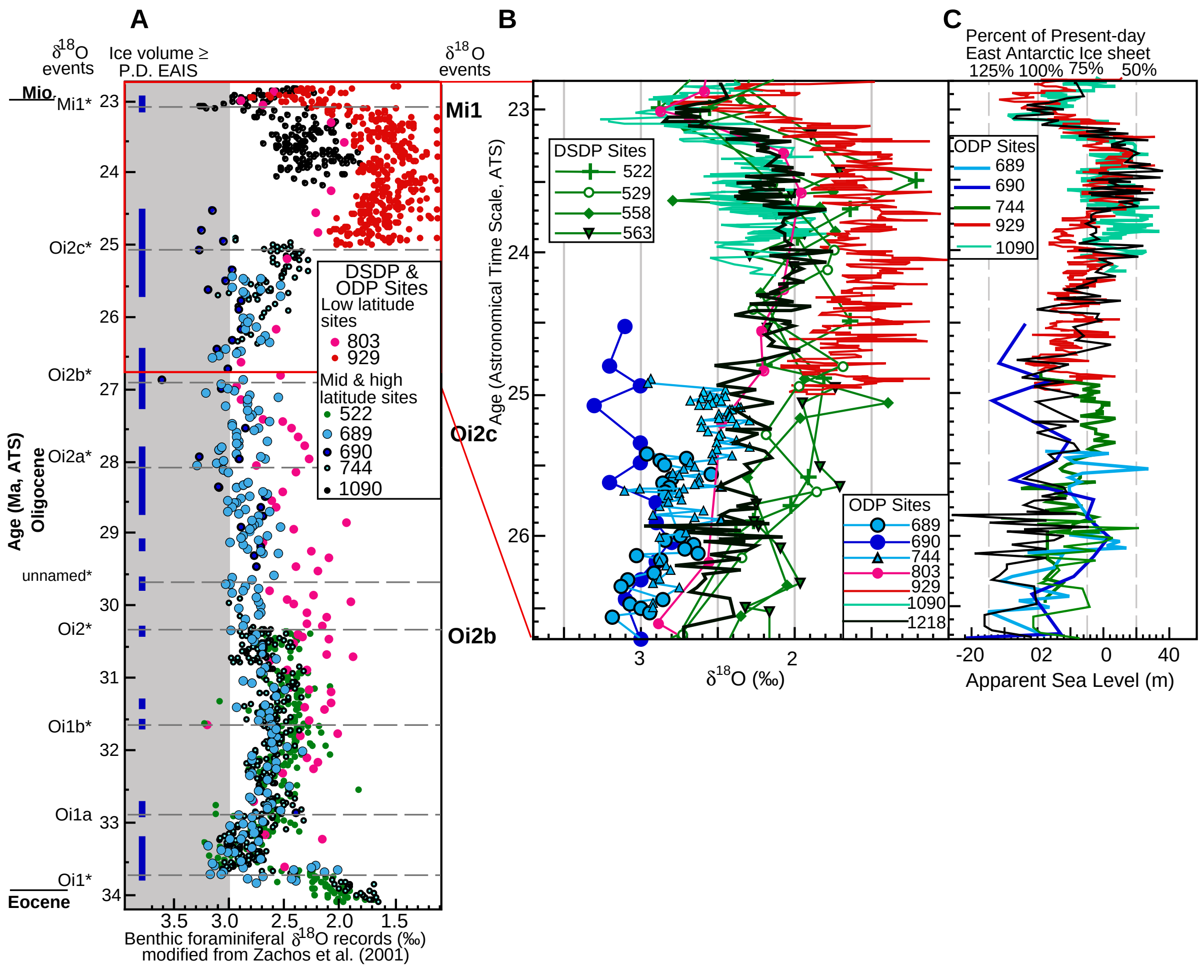

Fig. 3. (A) Oligocene benthic foraminiferal δ^{18}O records are from Barrera et al. (1993) and modified from Zachos et al. (2001). Blue circles are from Southern Ocean sites, green circles are from mid-latitude sites, and pink and red circles are from low latitude sites. The abrupt δ^{18}O decrease at circa 25.5 Ma is due to a change in the source of data from high latitude to low latitude sites, with Southern Ocean sites below and mainly western equatorial Atlantic Site 929 above (Pekar et al., 2002). (B) Oxygen isotope records for the late Oligocene showing the divergence in the records (from Barrera et al., 1993; modified from Zachos et al., 2001; Billups et al., 2002; Lear et al., 2004). Isotopic values of ≥ 3‰ are consistent with carbonate that formed in water of 2.0 °C concomitant with a fully glaciated Antarctic continent during the late Oligocene. Oxygen isotope values from *Cibicidoides* spp. are depleted relative to isotope equilibrium and were adjusted accordingly by 0.64‰ (Graham et al., 1981). On the left are δ^{18}O events from Miller et al. (1991) and Pekar and Miller (1996). (C) Apparent sea-level (ASL) estimates are derived from application of δ^{18}O to ASL calibrations to δ^{18}O records from Sites 689, 690, 744, 929, 1090, and 1218. The upper *x*-axis is the percent of the present-day EAIS (equivalent to ~60 m ASL). The lower *x*-axis is apparent sea-level change, with zero representing sea level resulting from ice volume equivalent to the present-day EAIS, with increasing values representing sea-level rise and negative numbers representing ice volume greater than the present-day EAIS volume.

563 (2.3‰ and 1.7‰), with the highest gradient (1.0‰ to 2.0‰) occurring between Site 690 and Site 929 between 25.0 and 24.5 Ma. Therefore, most of the abrupt late Oligocene δ^{18}O shift at ~25 Ma in the composite record of Zachos et al. (2001) is due to a change in the location of the data source, with values younger than 25 Ma being mainly from Atlantic Ocean Site 929 and values older than 25 Ma being from mid-latitude and Southern Ocean sites (Pekar et al., 2002; Lear et al., 2004). Additionally, a significant gradient also develops between Sites 690 and the only high-resolution record that extends throughout the late Oligocene, tropical Pacific Ocean Site 1218, ranging from ~0.5‰ to 1.2‰ near the top of the record at Site 690. Furthermore, no significant changes in isotopic values occur in the Site 1218 record above the top of the Site 690 record (24.5 Ma), supporting the idea that a large gradient (> 1‰) developed between the Weddell Sea (i.e., Site 690) and sites located in the Atlantic Basin. In summary, while aliasing is always a factor when com-

paring low-resolution records, it is clear that a significant $\delta^{18}O$ gradient developed during the late Oligocene between the Weddell Sea Sites and from the Atlantic (from 1.0‰ up to 2.0‰, equivalent to 4–8 °C) and to a lesser extent from the Pacific Ocean (~0.6‰ to ~1.0‰, equivalent to ~2–4 °C). This gradient is at least partly responsible for the significant decrease seen in other composite records (Miller et al., 1987; Abreu and Anderson, 1998), owing to that these composite records were biased to isotopic records from Atlantic Ocean Sites. These large $\delta^{18}O$ gradients allowed an evaluation of changing deep-sea water masses through the late Oligocene. In contrast, identifying deep-water masses using $\delta^{13}C$ records during the late Oligocene and early Miocene is difficult because basin-to-basin gradients were small (e.g., Woodruff and Savin, 1989; Wright et al., 1992; Wright and Miller, 1993). These small gradients have been ascribed to low mean ocean nutrient levels (Billups et al., 2002).

4. Deep-sea water mass distribution changes during the late Oligocene

Construction of three $\delta^{18}O$ transects for Oi-events Oi2b (27.0–26.6 Ma), Oi2c (25.2–24.8 Ma), and Mi1 (23.2–22.8 Ma) indicates that a significant water mass redistribution occurred during the late Oligocene (Fig. 4). A 400 ky time slice is used here to ensure that the maximum isotopic value of the $\delta^{18}O$ event is captured. The maximum $\delta^{18}O$ value from each site is used for the time slices as it should be the most representative of the $\delta^{18}O$ event. Although, low-resolution records often may not capture the highest value at an isotopic event, these isotopic transects do provide an indication of the broad changes that occurred at each event. Each of the time slices contains isotopic values of $\geq$ 3.0‰, which are consistent with carbonate forming in cold bottom-water ($\leq$ 2.0 °C) concomitant with a fully glaciated EAC. At 26.8 Ma, the highest $\delta^{18}O$ values are found at the Weddell Sea Sites (> 3‰), with slightly lower values in the Pacific and Indian basins (2.6‰–2.9‰) and Atlantic basin (2.3‰–3.0‰). Using $\delta^{18}O$ to ASL calibrations, a bottom-water temperature of 1.0–2.0 °C is estimated for the Weddell Sea (based on $\delta^{18}O$ values 3.0‰–3.6‰), with 2–4 °C in the Pacific and Indian Oceans and 2–5 °C at the Atlantic Sites. In contrast, during the Oi2c event, $\delta^{18}O$ values in the Southern Ocean (i.e., Sites 689, 690, and 744) remain high (2.9‰–3.3‰), while values in the Atlantic basin decreased by 0.5‰ to 1.0‰ (Fig. 4). This results in an isotopic gradient between the Southern and Atlantic Oceans of 1.0‰ to 1.2‰ (equivalent to ~4 to 5 °C, if the entire increase was due to temperature). This suggests that a second deep-water mass developed during the late Oligocene and replaced the colder deep-water in much of the ocean basins as suggested by the heavy isotopic values observed at Oi-event Oi2b. However, warmer water without increased salinity would have insufficient density to become a deep-water mass and compete with a colder dense water mass originating from Antarctica. If a bottom-water mass near Antarctic during the Oligocene were ~2 °C, with similar salinities as the present-day Antarctic Bottom Water (34.65‰; Wright and Colling, 1995), a bottom-water with a temperature of ~7 °C would require a salinity ~0.6‰ higher to have a similar density as the deep-water near Antarctica (based on Wright and Colling, 1995). This would result in an isotopic increase of ~0.2‰ (using 1.0‰ salinity = 0.3‰ $\delta^{18}O$), which would offset the temperature increase by 0.8 °C. Therefore, a large portion of deep-water temperatures in the Atlantic basin could have been ~7 °C or greater, based on a bottom-water temperature of 2.0 °C (based on values of 2.9‰ and 3.2‰) in the deep-water at Sites 690 and 744. These bottom-water temperatures are warmer than at any other time during the Oligocene and are similar to temperatures estimated during the middle Eocene (e.g., Miller et al., 1987; Zachos et al., 2001). It should also be noted that $\delta^{18}O$ values from Site 744 decrease slightly (2.7‰–2.3‰) and diverge by ~0.5‰ (equivalent to ~2 °C) from records from the Weddell Sea between 25.6 and 25.1 Ma before returning to similarly high values (~3‰) at 25.0 Ma. This suggests that slightly warmer deep-water may also have briefly bathed this region of the Southern Ocean during this time. During the Mi1 event, increasing $\delta^{18}O$ values indicate colder water once again filled most of the Atlantic basin, with the highest values (3.0‰–3.3‰) (and presumably coldest water) being found in the deepest part of the basin at Southern Atlantic Site 1090 and Southern Indian Ocean Site 704 (Fig. 4). Hiatuses at Sites 689, 690, and 744 prevented us from evaluating the spatial extent of colder Southern Component Water during this time interval. In contrast to the Oi2c event, warmer water at the Oligocene/Miocene boundary became restricted to intermediate water depths in the Indian and Atlantic Oceans.

5. Estimates of ice volume changes during the late Oligocene

Late Oligocene (26–23 Ma) ice volume estimates are equivalent to ~50% to 125% of present-day EAIS,

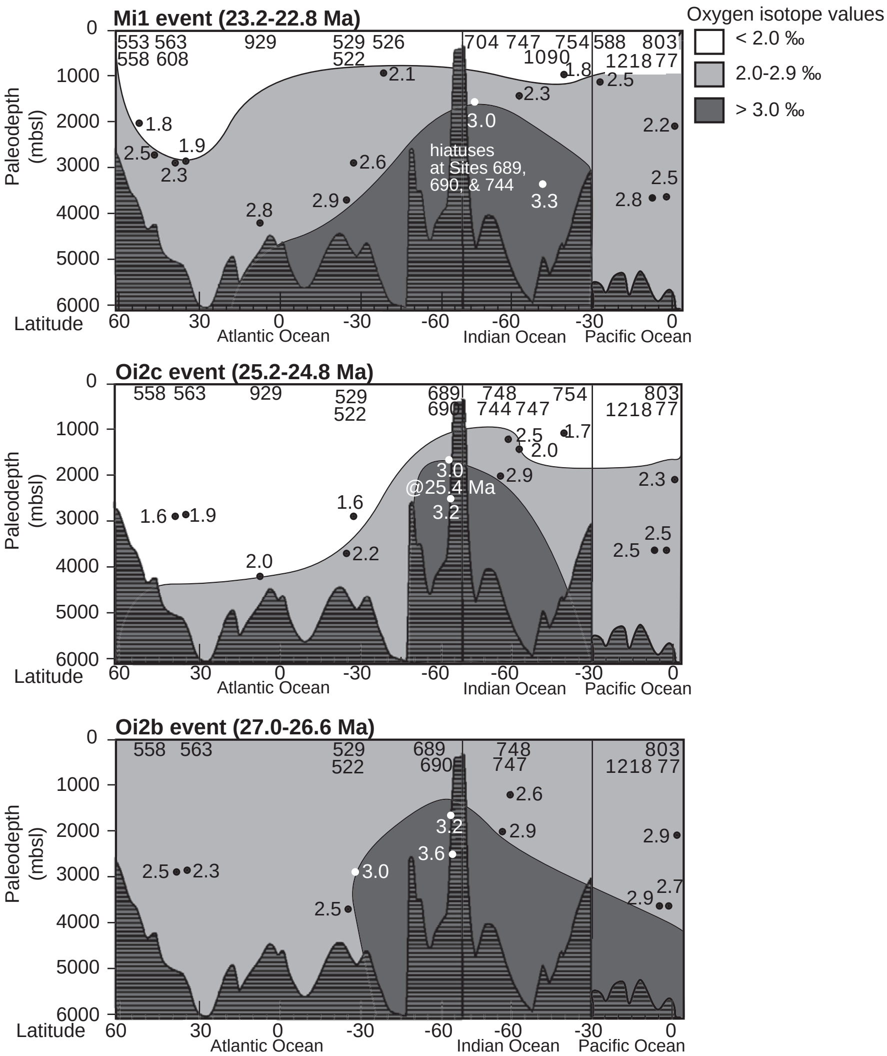

Fig. 4. Oxygen isotope transects for the Atlantic Ocean, Indian Ocean, and Pacific Oceans are constructed at three oxygen isotope events: (A) Oi2b (26.8 Ma), (B) Oi2c (25.1 Ma), and (C) Mi1 (23.0 Ma). DSDP and ODP sites used for each transect are labeled in the upper portion of each transect. Paleodepths for each site are indicated next to circles. Oxygen isotopic values shown next to the circles represent the maximum values within the 400-ky time slice. In each of the isotopic transects, cold deep-water (≥ 3.0‰) is generally restricted to the Southern Ocean. In contrast, warmer water (< 2.0‰) becomes an important water mass in the Atlantic and Indian Oceans during Oi-event Oi2c (circa 25.1 Ma). Note that the top of the record from Site 689 ends ~200 ky below the time slice for Oi2c event.

during glacial minima and maxima, respectively. High δ^{18}O values (2.9‰–3.3‰) consistent with a heavily glaciated EAC occurred at the top of the records at all three Southern Ocean sites (ODP Sites 689, 690, and 744) between 25.4 and 24.5 Ma (Fig. 3B). Between 25.0 and 24.5 Ma, an offset of ~1.2‰ exists between average δ^{18}O values from Site 690 and maximum values from Site 929. The new chronology by Florindo and Roberts (2005) provides a much higher confidence that the average values at Site 690 between 25.0 and 24.5 Ma correlate to the maximum values at Site 929, therefore permitting ice volume during the latest Oligocene to be constrained using the calibrated record from Site 929. A 1.0‰ decrease occurs between the high (glacial periods) and lowest δ^{18}O values (interglacial periods) from Site 929 from 25.0 to 24.0 Ma, which is equivalent to an ice volume decrease of 50% of the present-day EAIS. During the Mi1 event, the record from Site 1090 contains values (3.3‰; Billups et al., 2002) consistent with a return to a fully glaciated EAC, being perhaps 15% larger than the present-day EAIS. Therefore, increasing δ^{18}O values at Sites 929 and 1218 between 23.5 and 23.0 Ma are attributed to mainly bottom-water cooling concomitant with an increase in ice volume equivalent to ~40 m ASL fall.

The resolution of the $\delta^{18}O$ record from Site 929 (Zachos et al., 2001) is sufficiently high to resolve cycles that have been attributed to the 41 ky obliquity cycle. This permitted evaluation of Antarctic ice volume variability at the 10^4 yr timescale. Variability in the $\delta^{18}O$ record attributed to obliquity-forcing suggests changes in ice volume from 80% of the modern EAIS during glacial periods to 50% during interglacial periods. This range in variability is equivalent to ~20 m change in ASL and is generally consistent with orbitally forced ice sheet simulations under higher than present atmospheric CO_2 (DeConto and Pollard, 2003b).

6. Resolving the conundrum of warm deep-water concomitant with a glaciated Antarctica

The conundrum of low $\delta^{18}O$ values and a large ice sheet on Antarctica can be resolved by inferring the existence of at least two deep-water masses during the late Oligocene, one originating from near Antarctica and typically described as proto-Antarctic Bottom Water (proto-AABW) and a second warmer (and presumably more saline) water mass that strengthened after 27 Ma. It appears from $\delta^{18}O$ records from the Weddell Sea and other ocean basins that most of the time the colder proto-AABW was not the dominant deep-water mass outside the Weddell Sea. This suggests that this deep-water became entrained into and mixed with the warmer deep-water mass as it moved away from the Weddell Sea. The only exceptions occurred during Oi-events, Oi2b and Mi1, when colder deep-water filled the deep ocean basins (Fig. 4). During these glacial events, the expansion of the Antarctic ice sheet near the coastlines would have contributed to colder surface water temperatures and expanded sea ice cover, which likely impacted deep-water formation around Antarctica. This could have resulted in the strengthening of a proto-AABW during periods when the ice was at its maximum. In contrast, during glacial minima, retreat of the ice sheet into the continental interior would have resulted in a reduction of sea ice and warmer water temperatures along the coastline, leading to a reduction in proto-AABW production. Furthermore, additional runoff from a retreating ice sheet during interglacials would result in a freshening of the coastal water around Antarctic, further weakening proto-AABW production. This suggests that the large temperature gradient among deep-sea sites during the late Oligocene was the result of decreased production of proto-AABW and perhaps an increase in the production of warmer deep-water, and not a cooling of proto-AABW as a result of the opening of the Drake Passage and subsequent isolation of the Southern Ocean as previously suggested (Billups et al., 2002).

A number of numerical modeling simulation studies that tested the effects of open versus closed Southern Ocean gateways on ocean circulation and climate have yielded results that support the idea that as the Drake Passage opened, Southern Ocean deep-water formation (i.e., proto-AABW) decreased, high southern latitude temperatures decreased, and a warmer deep-water mass developed in the northern hemisphere (i.e., proto-North Atlantic Deep Water, proto-NADW) (Mikolajewicz et al., 1993; Toggweiler and Samuels, 1995; Nong et al., 2000; Toggweiler and Bjornsson, 2000; Huber et al., 2004; Sijp and England, 2004). The Drake Passage was likely the final barrier to circum-Antarctic circulation, opening in the early Oligocene (Lawver and Gahagan, 1998), and began providing a deep-water passage somewhat later in the Oligocene (Livermore et al., 2004). Sijp and England (2004), using an ocean general circulation model coupled to a simple climate model, showed that while high Southern Latitude SSTs cooled by several degrees when the Drake Passage opened, Southern Hemisphere deep-water formation (i.e., proto-AABW) slowed by ~75%, with little or no NADW until the passage opened, with significant NADW production beginning only after a deep circum-Antarctic passage was established. The simulations indicate that the turning on of NADW warmed the intermediate and deep ocean north of 30° South by ~2 °C. In the model, North Atlantic sea surface salinity also increased by ~1 psu, mostly in response to ocean circulation, rather than fresh water forcing at the surface. These results support the conclusions in this paper of reduced proto-AABW formation, strengthening of a warmer deep-water mass, and a cold and heavily glaciated East Antarctic continent occurring during the late Oligocene.

The Tethys has been suggested as a possible source of warmer water during the early Miocene (e.g., Brass et al., 1982; Woodruff and Savin, 1989). Isotopic evidence suggests that during the early Miocene, warmer deep to intermediate water formed in the Tethys and then entered the Indian Ocean (Woodruff and Savin, 1989; Wright and Miller, 1993). Warmer deep-water inferred by the low $\delta^{18}O$ values from Atlantic Ocean sites during the late Oligocene may represent the initiation of a warmer (and presumably more saline) deep-water mass, which may be analogous to the warmer water postulated to have originated from the Tethys in the early Miocene by Woodruff and Savin (1989). However, during the Oligocene, warm water appears to have originated in the Atlantic and not the Indian Ocean, as in the early Miocene.

Furthermore, high δ^{18}O values from Weddell Sea sites (3.0‰–3.3‰) occurred at times when most ocean basins contained δ^{18}O values that were similar to the low early Miocene δ^{18}O values. This suggests that the late Oligocene model of a spatially restricted proto-AABW, an expanded warmer deep-water mass, and a heavily glaciated East Antarctica may have also been true for the early Miocene (Pekar and DeConto, this issue). This may explain the evidence for cold, tundra-like conditions on the Ross Sea margin and continental glaciation, as inferred from proximal Antarctic records during this time interval (Raine, 1998; Cape Roberts Science Team, 1998; Roberts et al., 2003) as well as ice grounding lines across the Antarctic shelf in areas such as the Ross Sea and Prydz Bay (Cooper et al., 1991; Bartek et al., 1997). This contrasts with previous interpretations of Antarctic warming during the early Miocene based on δ^{18}O records and is also consistent with recent numerical modeling studies (DeConto and Pollard, 2003a,b), which showed how significant Antarctic ice can exist during times with warmer-than-present-day global mean temperatures, and poleward ocean heat transport.

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References

Abreu, V.S., Anderson, J.B., 1998. Glacial eustasy during the Cenozoic: sequence stratigraphic implications. American Association of Petroleum Geologists Bulletin 82, 1385–1400.

Axelrod, D.I., Raven, P.H., 1985. Origins of the Cordilleran flora. Journal of Biogeography 12, 21–47.

Barrera, E., Baldauf, J., Lohmann, K.C., 1993. Strontium isotope and benthic foraminifer stable isotopic results from Oligocene sediments at Site 803. In: Berger, W.H., Kroenke, L.W., Mayer, L.A. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results, Leg, vol. 130. Ocean Drilling Program, College Station, Texas, pp. 269–279.

Barrett, P.J. (Ed.), 1989. Antarctic Cenozoic history from CIROS-1 drillhole, McMurdo Sound, DSIR Bulletin, vol. 245. Scientific Information Publishing Centre, Wellington (254 pp.)

Bartek, L.R., Andersen, J.L.R., Oneacre, T.A., 1997. Substrate control on distribution of subglacial and galciomarine seismic facies based on stochastic models of glacial seismic facies deposition on the Ross Sea continental margin, Antarctica. Marine Geology 143, 223–262.

Billups, K., Channell, J.E.T., Zachos, J., 2002. Late Oligocene to early Miocene geochronology and paleoceanography from the subantarctic South Atlantic. Paleoceanography 17, doi:10.1029/2000PA000568.

Brass, G.W., Southam, J.R., Peterson, W.H., 1982. Warm saline bottom water in the ancient ocean. Nature 296, 620–623.

Cape Roberts Science Team, 1998. Initial Report on CRP-1, Cape Roberts Project, Antarctica. Terra Antartica 5 (187 pp.).

Cape Roberts Science Team, 1999. Studies from the Cape Roberts Project, Ross Sea, Antarctica, Initial Report on CRP-2/2A. Terra Antartica 6 (187 pp.).

Cape Roberts Science Team, 2000. Studies from the Cape Roberts Project, Ross Sea, Antarctica, Initial Report on CRP-3. Terra Antartica 7 (209 pp.).

Cooper, A.K., Barrett, P.J., Hinz, K., Traube, V., Leitchenkov, G., Stagg, H.M.J., 1991. Cenozoic prograding sequences of the Antarctic continental margin: a record of glacio-eustatic and tectonic events. Marine Geology 102, 175–213.

Coxall, H.K., Wilson, P.A., Palike, H., Lear, C.H., Backman, J., 2005. Rapid stepwise onset of Antarctic glaciation and deeper calcite compensation in the Pacific Ocean. Nature 433, 53–57.

DeConto, R.M., Pollard, D., 2003a. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. Nature 421, 245–249.

DeConto, R.M., Pollard, D., 2003b. A coupled climate-ice sheet modeling approach to the early Cenozoic history of the Antarctic ice sheet. Palaeogeography, Palaeoclimatology, Palaeoecology 198, 39–52.

Denton, G.H., Hughes, T.J., 2002. Reconstructing the Antarctic ice sheet at the last glacial maximum. Quaternary Science Reviews 21, 193–202.

Florindo, F., Roberts, A.P., 2005. Eocene–Oligocene magnetobiochronology of ODP Sites 689 and 690, Maud Rise, Weddell Sea, Antarctica. Geological Society of America Bulletin 117, 46–66.

Graham, D.W., Corliss, B.H., Bender, M.L., Keigwin, L.D., 1981. Carbon and oxygen isotopic disequilibria of recent benthic foraminifera. Marine Micropaleontology 6, 483–497.

Hay, W.W., DeConto, R., Wold, C.N., Wilson, K.M., Voigt, S., Schulz, M., Wold-Rossby, A., Dullo, W.-C., Ronov, A.B., Balukhovsky, A.N., Soeding, E., 1999. Alternative global Cretaceous paleogeography. In: Barrera, E., Johnson, C. (Eds.), The Evolution of Cretaceous Ocean/Climate Systems. Geological Society of America Special Paper, vol. 332, pp. 1–47 (http://www.odsn.de/odsn/services/paleomap/paleomap.html).

Huber, M., Brinkhuis, H., Stickley, C.E., Doos, L., Sluijs, A., Warnaar, J., Schellenberg, S.A., Williams, G.L., 2004. Eocene circulation of the Southern Ocean: was Antarctica kept warm by subtropical waters? Paleoceanography 19, A4026.

Huybrechts, P., 2002. Sea-level changes at the LGM from ice-dynamic reconstructions of the Greenland and Antarctic ice sheets during the glacial cycles. Quaternary Science Reviews 21, 203–231.

Kennett, J.P., Stott, L.D., 1990. Proteus and proto-Oceanus: Paleogene oceans as revealed from Antarctic stable isotopic results. ODP Leg 113. Proceedings of the Ocean Drilling Program Scientific Results 94, 1217–1244.

Kominz, M.A., Pekar, S.F., 2001. Oligocene eustasy from two-dimensional sequence stratigraphic backstripping. Geological Society of America Bulletin 113, 291–304.

Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A., Levrard, B., 2005. A long term numerical solution for the insolation quantities of the Earth. Astronomy and Astrophysics 428, 261–286, doi:10.1051/0004-6361:20041335.

Lawver, L.A., Gahagan, L.M., 1998. Opening of Drake passage and its impact on Cenozoic ocean circulation. In: Crowley, T.J., Burke, K.C. (Eds.), Tectonic Boundary Conditions for Climate Reconstructions. Oxford University Press, New York, pp. 212–223.

Lear, C.H., Elderfield, H., Wilson, P.A., 2000. Cenozoic deep-sea temperature and global ice volumes from Mg/Ca in benthic foraminiferal calcite. Science 287, 269–272.

Lear, C.H., Rosenthal, Y., Coxall, H.K., Wilson, P.A., 2004. Late Eocene to early Miocene ice sheet dynamics and the global carbon cycle. Paleoceanography 19, A4015, doi:10.1029/2004PA001039.

Livermore, R., Eagles, G., Morris, P., Maldonado, A., 2004. Shackleton fracture zone: no barrier to early circumpolar ocean circulation. Geology 32, 797–800.

Mikolajewicz, U., Maier-Reimer, E., Crowley, T.J., Kim, K.-Y., 1993. Effect of Drake and Panamanian gateways on the circulation of an ocean model. Paleoceanography 8 (4), 409–426.

Miller, K.G., Fairbanks, R.G., 1983. Evidence for Oligocene-Middle Miocene abyssal circulation changes in the western North Atlantic. Nature 306, 250–253.

Miller, K.G., Thomas, E., 1985. Late Eocene to Oligocene benthic isotopic record, Site 574, Equatorial Pacific. Initial Report Deep Sea Drilling Project 85, 981–996.

Miller, K.G., Fairbanks, R.G., Mountain, G.S., 1987. Tertiary oxygen isotope synthesis, sea-level history, and continental margin erosion. Paleoceanography 2, 1–19.

Miller, K.G., Feigenson, M.D., Kent, D.V., Olsson, R.K., 1988. Oligocene stable isotope ($^{87}Sr/^{86}Sr$, $\delta^{18}O$, $\delta^{13}C$) standard section. Deep Sea Drilling Project. Paleoceanography 3, 223–233.

Miller, K.G., Wright, J.D., Fairbanks, R.G., 1991. Unlocking the icehouse: Oligocene–Miocene oxygen isotopes, eustasy, and margin erosion. Journal of Geophysical Research 96, 6,829–6,848.

Naish, T.R., Wolfe, K.J., Barrett, P.J., Wilson, G.S., Cliff, A., Bohaty, S.M., Bucker, C.J., Claps, M., Davey, F.J., Dunbar, G.B., Dunn, A.G., Fielding, C.R., Florindo, F., Hannah, M.J., Harwood, D.M., Henrys, S.A., Krissek, L.A., Lavelle, M., van der Meer, J., McIntosh, W.C., Niessen, F., Passchier, S., Powell, R.D., Roberts, A.P., Sagnotti, L., Scherer, R.P., Strong, P.C., Talarico, F., Verosub, K.L., Villa, G., Watkins, D.K., Webb, P.-N., Wonik, T., 2001. Orbitally induced oscillations in the East Antarctic ice sheet at the Oligocene/Miocene boundary. Nature 413, 719–723.

Nong, G.T., Najjar, R.G., Seidov, D., Peterson, W., 2000. Simulation of ocean temperature change due to the opening of Drake Passage. Geophysical Research Letters 27, 2689–2692.

Pekar, S.F., DeConto, R., this issue. High-Resolution Ice-Volume Estimates for the Early Miocene: Evidence for a Dynamic Ice Sheet in Antarctica, Palaeogeography, Palaeoclimatology, Palaeoecology.

Pekar, S.F., Kominz, M.A., 2001. Two-dimensional paleoslope modeling: a new method for estimating water depths for benthic foraminiferal biofacies and paleo shelf margins. Journal of Sedimentary Research 71, 608–620.

Pekar, S., Miller, K.G., 1996. New Jersey Oligocene “Icehouse” sequences (ODP Leg 150X) correlated with global $\delta^{18}O$ and Exxon eustatic records. Geology 24, 567–570.

Pekar, S.F., Christie-Blick, N., Kominz, M.A., Miller, K.G., 2002. Calibrating eustasy to oxygen isotopes for the early icehouse world of the Oligocene. Geology 30, 903–906.

Prebble, J.G., Raine, J.I., Barrett, P.J., Hannah, M.J., this issue. Vegetation and climate from two Oligocene glacioeustatic sedimentary cycles (31 and 24 Ma) cored by the Cape Roberts Project, Victoria Land Basin, Antarctica. Palaeogeography, Palaeoclimatology, Palaeoecology.

Raine, J.I., 1998. Terrestrial palynomorphs from Cape Roberts Project drillhole CRP-1, Ross Sea, Antarctica. Terra Antarctica 5, 539–548.

Raine, J.I., Askin, R.A., 2001. Oligocene and Early Miocene terrestrial palynology of the Cape Roberts Drillhole CRP-2/2A, Victoria Land Basin, Antarctica. Terra Antarctica 7, 389–400.

Ritz, C.V., Rommelaere, V., Dumas, C., 2001. Modeling the evolution of Antarctic ice sheet over the last 420,000 years: implications for altitude changes in the Vostok region. Journal of Geophysical Research 106 (D23), 31943–31964.

Roberts, A.P., Wilson, G.S., Harwood, D.M., Verosub, K.L., 2003. Glaciation across the Oligocene–Miocene boundary in Southern McMurdo Sound, Antarctica: new chronology from the CIROS-1 drill hole. Palaeogeography, Paleoclimatology, Palaeoecology 198, 113–130.

Shackleton, N.J., Kennett, J.P., 1975. Late Cenozoic oxygen and carbon isotopic changes at DSDP Site 284: implications for glacial history of the Northern hemisphere and Antarctic. In: Kennett, J.P., Houtz, R.E., et al., (Eds.), Initial Reports of the Deep Sea Drilling Project, vol. 29, pp. 801–808.

Sijp, W.P., England, M.H., 2004. Effect of the Drake Passage through flow on Global Climate. Journal of Physical Oceanography 34, 1254–1266.

Speiss, V., 1990. Cenozoic Magnetostratigraphy of Leg 113 drill sites, Maude Rise, Weddell Sea, Antarctica. In: Barker, P.F., Kennett, J.P. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results, vol. 113, pp. 261–315.

Tiffney, B.H., 1985. The Eocene North Atlantic land bridge: its importance in Tertiary and modern phytogeography of the Northern Hemisphere. Journal of the Arnold Arboretum 66, 243–273.

Thomas, E., Zahn, R., Diester-Hauss, L., 1995. The Eocene–Oligocene transition at high latitudes: benthic foraminifera, sediments and stable isotopes. EOS Transactions 76 (Supplement), S187.

Thorn, V.C., 2001. Oligocene and Early Miocene phytoliths from CRP-2/2A and CRP-3, Victoria Land Basin, Antarctica. Terra Antarctica 8, 407–422.

Toggweiler, J.R., Bjornsson, H., 2000. Drake Passage and paleoclimate. Journal of Quaternary Science 15, 319–328.

Toggweiler, J.R., Samuels, B., 1995. Effect of Drake Passage on the global thermohaline circulation. Deep-Sea Research 42, 477–500.

Williams, R.S., Ferrigno, J.G., 1999. Satellite Image Atlas of Glaciers of the World, Chapter A: Introduction. U.S. Geological Survey Professional Paper 1386-A.

Wolfe, J.A., 1978. A paleobotanical interpretation of Tertiary climates in the Northern Hemisphere. American Scientist 66, 694–704.

Wolfe, J.A., Poore, R.Z., 1982. Tertiary marine and nonmarine climatic trends. In: Crowell, J.C., Berger, W. (Eds.), Pre-Pleistocene Climates, pp. 154–158 (Washington).

Woodruff, F., Savin, S.M., 1989. Miocene deep-water oceanography. Paleoceanography 4, 87–140.

Wright, J., Colling, A. 1995. Seawater: Its Composition, Properties, and Behaviour. Open University Press and Elsevier Science Ltd., Oxford, 168.

Wright, J.D., Miller, K.G., 1993. Southern ocean influences on late Eocene to Miocene deep water circulation. American Geophysical Union, Antarctic Research Series 60, 1–25.

Wright, J.D., Miller, K.G., Fairbanks, R.G., 1992. Early and middle Miocene stable isotopes: implication for deep-water circulation and climate. Paleoceanography 7, 357–387.

Zachos, J.C., Berggren, W.A., Aubry, M.P., Mackensen, A., 1992. Isotope and trace element geochemistry of Eocene and Oligocene Foraminifers from Site 748, Kerguelen Plateau. In: Schlich, R., Wise Jr., S.W., et al. (Eds.), Proc. ODP, Scientific Results, vol. 120. Ocean Drilling Program, College Station, TX, pp. 839–854.

Zachos, J.C., Pagani, M., Sloan, L., Thomas, E., Billups, K., 2001. Trends, rhythms and aberrations in global climate 65 Ma to present. Science 292, 686–693.