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Climate threshold at the Eocene-Oligocene transition: Antarctic ice sheet influence on ocean circulation

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ABSTRACT

We present an overview of the Eocene-Oligocene transition from a marine perspective and posit that growth of a continent-scale Antarctic ice sheet ($25 \times 10^6 \text{ km}^3$) was a primary cause of a dramatic reorganization of ocean circulation and chemistry. The Eocene-Oligocene transition (EOT) was the culmination of long-term (10^7 yr scale) CO_2 drawdown and related cooling that triggered a 0.5‰–0.9‰ transient precursor benthic foraminiferal $\delta^{18}\text{O}$ increase at 33.80 Ma (EOT-1), a 0.8‰ $\delta^{18}\text{O}$ increase at 33.63 Ma (EOT-2), and a 1.0‰ $\delta^{18}\text{O}$ increase at 33.55 Ma (oxygen isotope event Oi-1). We show that a small (~ 25 m) sea-level lowering was associated with the precursor EOT-1 increase, suggesting that the $\delta^{18}\text{O}$ increase primarily reflected 1–2 °C of cooling. Global sea level dropped by 80 ± 25 m at Oi-1 time, implying that the deep-sea foraminiferal $\delta^{18}\text{O}$ increase was due to the growth of a continent-sized Antarctic ice sheet and 1–4 °C of cooling. The Antarctic ice sheet reached the coastline for the first time at ca. 33.6 Ma and became a driver of Antarctic circulation, which in turn affected global climate, causing increased latitudinal thermal gradients and a “spinning up” of the oceans that resulted in: (1) increased thermohaline circulation and erosional pulses of Northern Component Water and Antarctic Bottom Water; (2) increased deep-basin ventilation, which caused a decrease in oceanic residence time, a decrease in deep-ocean acidity, and a deepening of the calcite compensation depth (CCD); and (3) increased diatom diversity due to intensified upwelling.

Keywords: Antarctica, sea level, oxygen isotopes, Oligocene, Eocene.

INTRODUCTION: THE BIG CHILL

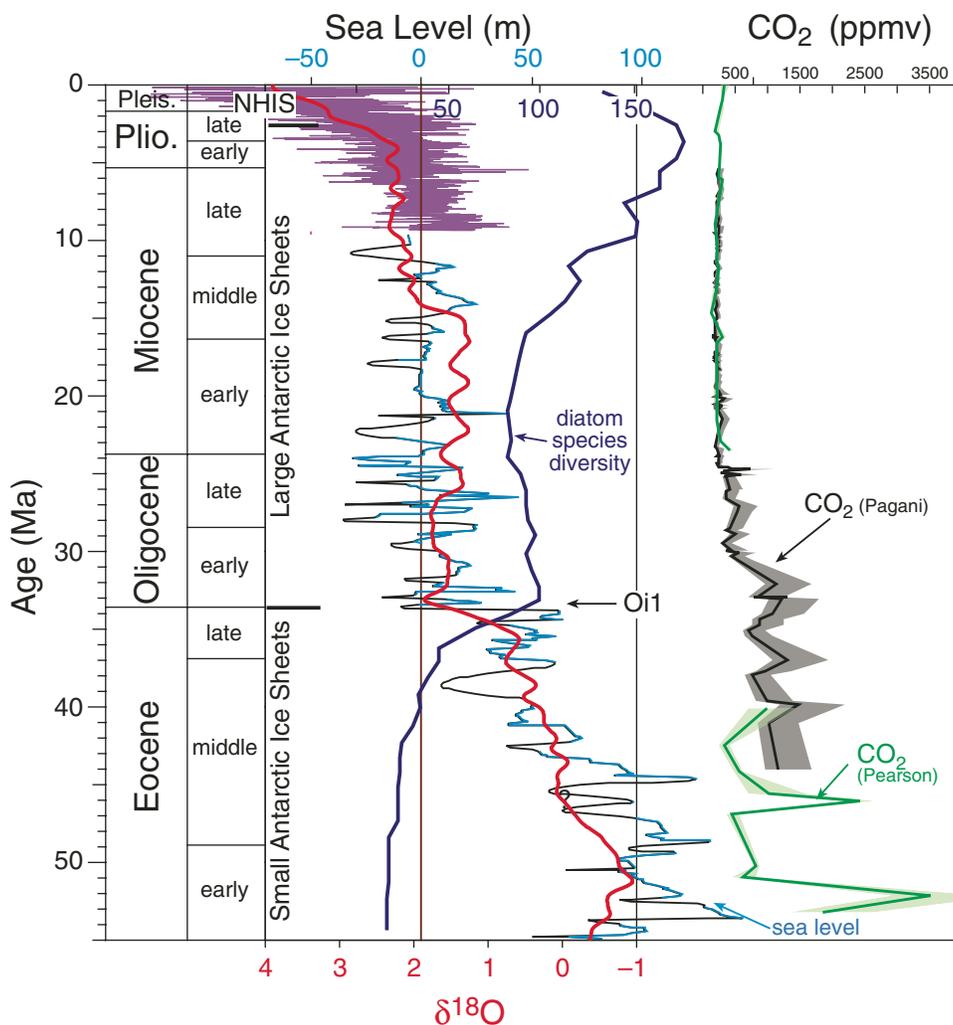
The Eocene-Oligocene transition is the most profound oceanographic and climatic change of the past 50 m.y. Cooling began in the middle Eocene and culminated in the major earliest Oligocene $\delta^{18}\text{O}$ increase (oxygen isotope event Oi-1, ca. 33.55 Ma). The Eocene-Oligocene transition is the largest of three Cenozoic deep-sea benthic foraminiferal $\delta^{18}\text{O}$ increases (Fig. 1), which also include the middle Miocene (ca. 14.8 Ma), when a permanent (i.e., dry-based) Antarctic ice sheet developed (Shackleton and Kennett, 1975), and the late Pliocene (ca. 2.6 Ma), when Northern Hemisphere ice volume increased (Shackleton et al., 1984) (“the Ice Ages”). Most studies agree that the earliest Oligocene $\delta^{18}\text{O}$ increase (Oi-1; 33.55 Ma) signaled the beginning of the icehouse Earth, with large ice sheets on Antarctica (Miller et al., 1991, 2005a, 2005b; Zachos et al., 1996). The 33.55 Ma event is marked by a 1.0‰ increase in deep-sea benthic foraminiferal $\delta^{18}\text{O}$ throughout the Atlantic, Pacific, Indian, and Southern Oceans (Corliss et al., 1984; Coxall et al., 2005; Keigwin, 1980; Kennett and Shackleton, 1976; Miller et al., 1987; Savin

et al., 1975; Shackleton and Kennett, 1975; Zachos et al., 2001) (Fig. 1). A latest Eocene precursor 0.5‰ $\delta^{18}\text{O}$ increase has been documented in the deep Pacific (Coxall et al., 2005) and Gulf Coast paleoshelf (Miller et al., 2008a; Katz et al., 2008), yielding an overall increase of ~1.5 ‰ across the Eocene-Oligocene transition (Fig. 2).

Deep-sea benthic foraminiferal $\delta^{18}\text{O}$ records reflect changes in both temperature and $\delta^{18}\text{O}_{\text{seawater}}$ due to ice-volume changes, complicating interpretations of the Eocene-Oligocene transition.¹ Over the past 50 m.y., long-term (10⁷ yr) deep-sea benthic foraminiferal $\delta^{18}\text{O}$ values have increased by ~5‰ (Miller et al., 1987, 2005a) (Fig. 1). Assuming that ice-volume changes can only explain ~2.0‰ of the total increase (assuming 0.11‰ per 10 m sea-level change; Fairbanks and Matthews, 1978; see following for discussion of this calibration), more than half of the long-term increase must be attributed to an ~12 °C cooling.

¹Though local fractionation resulting from evaporation/precipitation changes affects the surface ocean, it has a minimal effect on deep-sea $\delta^{18}\text{O}$ records, which reflect ice volume and deep-sea temperature (which in turn reflect high-latitude surface temperature).

Figure 1. Comparison of the global sea level (blue indicates intervals constrained by data; black indicates estimated lowstands; Kominz et al., 2008), benthic foraminiferal $\delta^{18}\text{O}$ values (reported to *Cibicidoides* and smoothed to remove periods shorter than 1 m.y.), diatom diversity curve (Katz et al., 2005), and atmospheric CO_2 estimates derived from alkenones (Pagani et al., 1999, 2005; stippled pattern—error) and boron isotopes (Pearson and Palmer, 2000; stippled pattern—error). NHIS—Northern Hemisphere ice sheets. The benthic foraminiferal curve is the Atlantic synthesis of Miller et al. (2005a; <http://geology.rutgers.edu/miller.shtml>), which is smoothed to remove periods shorter than 1 m.y.



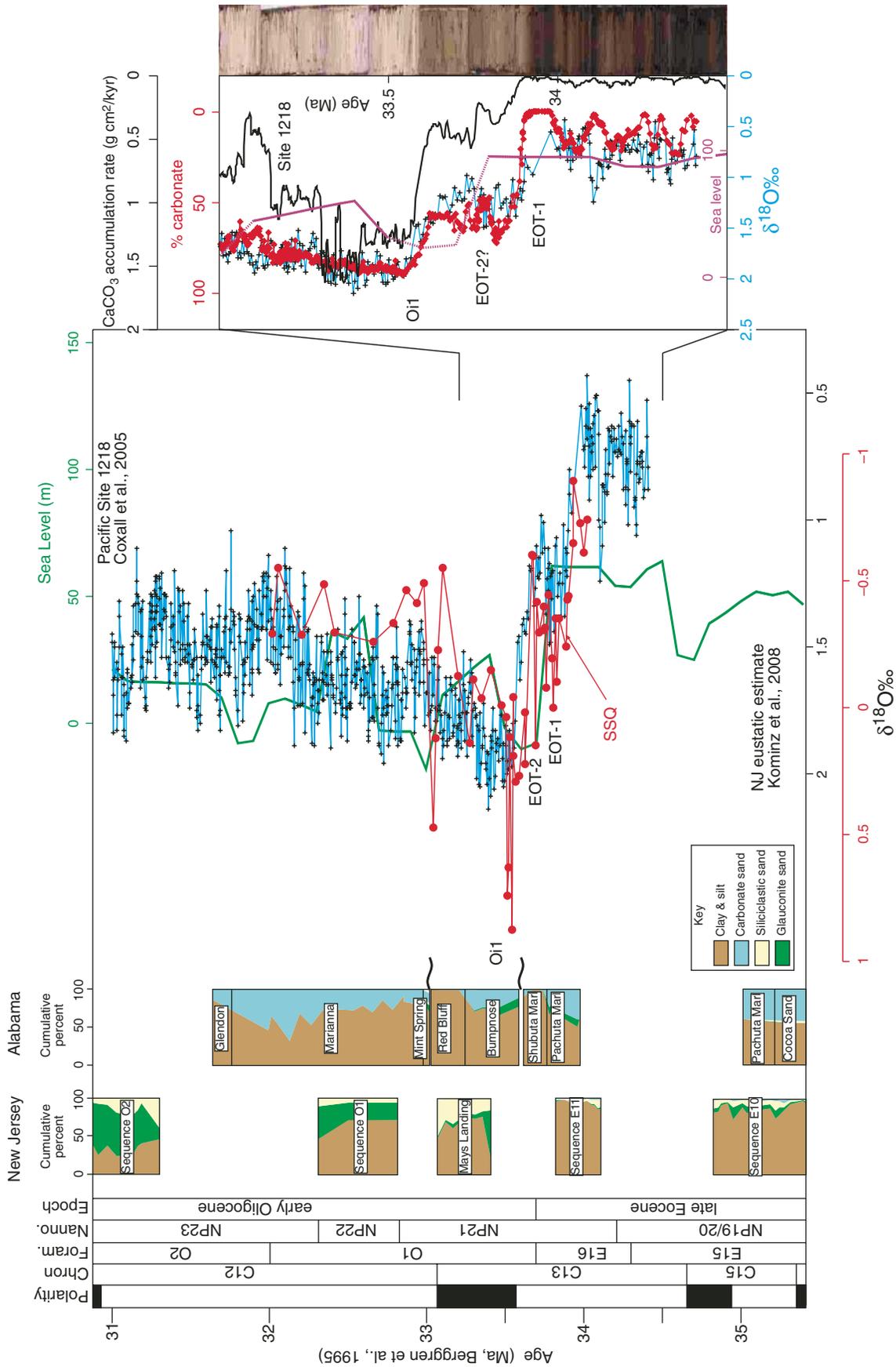


Figure 2. Comparison of continental margin records from New Jersey and Alabama (Miller et al., 2008a), global sea-level estimates (Kominz et al., 2008), and benthic foraminiferal $\delta^{18}\text{O}$ records from deep Pacific ODP Site 1218 (Coxall et al., 2005) and St. Stephens Quarry (SSQ), Alabama (Katz et al., 2008). The isotopes from SSQ provide first-order correlations to the sequences and sequence boundaries at SSQ. Close-up of Site 1218 oxygen isotopes at right includes carbonate percent and mass accumulation rate (AR) data (Coxall et al., 2005) and also shows two large drops associated with the EOT-1 and Oi-1 oxygen isotope increases, which are reflected in the core photograph at right (light color—carbonate rich, dark—carbonate poor). Figure was modified after Miller et al. (2008a).

Benthic foraminiferal Mg/Ca data show 12 °C of cooling over the last 50 m.y. (Lear et al., 2000), supporting this interpretation. On shorter time scales (10^5 – 10^6 yr), distinguishing the relative contributions of ice versus temperature on $\delta^{18}\text{O}$ records is difficult. As a result, considerable controversy has surrounded hypotheses about the cause(s) of the latest Eocene through earliest Oligocene $\delta^{18}\text{O}$ increases, ranging from early studies that attributed them entirely to deep-water cooling (and hence, high-latitude surface-water cooling; Shackleton and Kennett, 1975), to recent studies that ascribe the increase largely to ice-sheet expansion (Tripathi et al., 2005). This latter interpretation requires continental ice storage that is ~ 1.5 times larger than modern ice sheets in Antarctica and the Northern Hemisphere, and it predicts a 150 m global sea-level (eustatic) fall. The extent of bipolar glaciation has been disputed (Edgar et al., 2007; Katz et al., 2008; Miller et al., 2008a) and is discussed in detail later.

The Eocene-Oligocene transition was associated with a dramatic oceanographic reorganization and the largest climatic cooling of the Cenozoic. There was a major increase in ocean productivity (see summary in Berger, 2007), a very large drop in the calcite compensation depth (CCD) (Van Andel, 1975; Coxall et al., 2005; Rea and Lyle, 2005), and pulses of strongly eroding Antarctic bottom water (see summaries in Kennett [1977] and Wright and Miller [1996]) and Northern Component Water (see summaries in Tucholke and Mountain, 1979; Miller and Tucholke, 1983; and Miller, 1992). Paleontological evidence for cooling includes the development of psychrospheric (“cold-loving”) ostracods (Benson, 1975), deep-sea and shelf benthic foraminiferal extinctions and appearances (e.g., Miller et al., 1992; Thomas, 1992), extinctions in tropical planktonic and larger foraminifera (Adams et al., 1986; Keller et al., 1992; Pearson et al., 2008; Wade and Pearson, 2008), and the decline of thermophilic calcareous nannoplankton (Aubry, 1992; Dunkley Jones et al., 2008). Terrestrial cooling is indicated by pollen changes (e.g., New Jersey; Owens et al., 1988) and a mammalian turnover (e.g., England; Hooker et al., 2004). Regional aridification was associated with terrestrial cooling (e.g., the Himalayas; Dupont-Nivet et al., 2007; Zanazzi et al., 2007).

Various hypotheses linking these Eocene-Oligocene changes have been postulated. Until recently, the predominant hypothesis has been that the undocking of Australia from Antarctica and the opening of the Drake Passage led to development of the Antarctic Circumpolar Current. This thermally isolated Antarctica from the relatively warm water coming from the Subtropical Gyre, thereby cooling the continent (Kennett, 1977). The role of the Antarctic Circumpolar Current and thermal isolation has recently been both affirmed (Exon et al., 2004) and challenged (Huber et al., 2004), leaving the role of the gateway uncertain. Alternatively, a coupled global climate-ocean-ice model of the Eocene-Oligocene transition (DeConto and Pollard, 2003) has suggested that pervasive glaciation of Antarctica was triggered by a drop of atmospheric CO_2 below a critical threshold of 2.8 times pre-anthropogenic levels. The DeConto and Pollard (2003) model also suggested that opening of the Drake Passage could only have had a role in triggering the glaciation in the context of atmospheric CO_2

already near the critical threshold, although the increase in biological productivity, stimulated by the initiation/increase of the Antarctic Circumpolar Current, might have indirectly contributed to lowering atmospheric $p\text{CO}_2$ (Scher and Martin, 2006). Though the role of gateways versus CO_2 remains unresolved, it is clear that cooling in Antarctica resulted in the development of a large ice sheet across the Eocene-Oligocene transition. We hypothesize here that when the Antarctic ice sheet grew above a critical threshold, extending beyond the coastline for the first time, it began to influence ocean and climate changes.

Here, we synthesize published $\delta^{18}\text{O}$ records spanning the Eocene-Oligocene transition (Fig. 1) and compare them with records of sea level (Miller et al., 2005a; Kominz et al., 2008), passive margin sequences (Miller et al., 2005a, 2008a), CO_2 (Pagani et al., 1999, 2005; Pearson and Palmer, 2000), diatom diversity (Katz et al., 2005), and deep-sea carbonate accumulation (Coxall et al., 2005). We distinguish temperature from ice-volume effects by comparing $\delta^{18}\text{O}$ records (Figs. 1 and 2) with: (1) a global sea-level estimate that was derived from backstripping² New Jersey coastal plain core holes (Miller et al., 2005a; Kominz et al., 2008) and from evaluation of sequence stratigraphic records from St. Stephens Quarry (SSQ), Alabama (Miller et al., 2008a); and (2) by comparison with published Mg/Ca records from Deep Sea Drilling Project (DSDP) Site 522 (Lear et al., 2000), Ocean Drilling Program (ODP) Site 1218 (Lear et al., 2004), and SSQ (Katz et al., 2008).

ICE VOLUME VERSUS TEMPERATURE CHANGES AND THE EOCENE-OLIGOCENE TRANSITION

The Precursor (EOT1 and EOT2) Increases

Katz et al. (2008) named the transient precursor $\delta^{18}\text{O}$ increase (33.8 Ma) at Sites 1218 (Coxall et al., 2005), 522 (Zachos et al., 1996), and SSQ (Miller et al., 2008a; Katz et al., 2008) “EOT-1” for “Eocene-Oligocene transition event 1.” EOT-1 is associated with an extinction in planktonic foraminifera (the *Turborotalia cerroazulensis* lineage) (Miller et al., 2008a; Pearson et al., 2008) that has been dated at 33.80 Ma (Berggren et al., 1995). At SSQ, EOT-1 is manifested as a 0.9‰ $\delta^{18}\text{O}$ increase in neritic (~ 75 m paleodepth) benthic foraminifera and 2.5 °C cooling as determined by Mg/Ca paleothermometry (Katz et al., 2008). A similar cooling (2.0 °C) at deep-sea Sites 522 and 1218 determined by Mg/Ca is associated with benthic foraminiferal $\delta^{18}\text{O}$ increases of ~ 0.5 ‰. This suggests that a $\delta^{18}\text{O}_{\text{seawater}}$ increase of ~ 0.4 ‰ (SSQ) to 0.1‰ (522 and 1218) was associated with EOT-1 (Katz et al., 2008). A $\delta^{18}\text{O}_{\text{seawater}}$ change of 0.1‰–0.4‰ would have resulted from an ~ 10 – 35 m sea-level fall using calibrations discussed herein later.

²This is a method that progressively removes the effects of compaction, loading (either Airy or two-dimensional flexural loading), and thermal subsidence from margin records. The residual is the result of eustatic and nonthermal tectonic changes. In the absence of nonthermal tectonism, it provides a eustatic estimate.

Sea-level records across EOT-1 are not clear, in part because the published sea-level estimates (Miller et al., 2005a; Kominz et al., 2008) are not sensitive to small changes on short time scales (<1 m.y.). EOT-1 falls near a long hiatus in New Jersey, and, given the temporal resolution, it is likely that this hiatus is a concatenation of EOT-1 and Oi-1. Thus, the sea-level estimates based on New Jersey sequences (Miller et al., 2005a; Kominz et al., 2008) do not reflect this drop. Published sequence stratigraphic analyses at SSQ did not reveal a sea-level fall associated with the EOT-1 increase measured directly on the core (Miller et al., 2008a); however, a glauconite bed occurs at the same level as the EOT-1 increase at SSQ. This glauconite bed may represent a previously undetected unconformity and associated hiatus (i.e., glauconite sands often overlie sequence boundaries in the coastal plain because lowstand deposits are largely missing; Miller et al., 2005a). The SSQ sections are not sensitive to sea-level variations that are less than ~25 m because they were deposited in relatively deep shelf waters (~75 m; Miller et al., 2008a). As noted by Katz et al. (2008), a global sea-level fall at this time is supported by a correlative erosional event in the Priabonian type section (Brinkhuis and Visscher, 1995) in inner neritic carbonate facies most sensitive to small sea-level changes. We follow Katz et al. (2008) in interpreting EOT-1 as primarily a global cooling event of 2 °C and minor sea-level fall (~25 m).

Katz et al. (2008) noted a second earliest Oligocene (ca. 33.63 Ma) ~0.8‰ $\delta^{18}\text{O}$ increase, termed EOT-2, at SSQ (with no detectible hiatus), Site 522, and Site 1218. Mg/Ca data are lacking for this event, and its significance in terms of temperature versus ice volume remains unclear. The EOT-2 $\delta^{18}\text{O}$ increase occurs within a sequence at SSQ (within the Shubuta Marl) in a section with no marked lithologic or sequence stratigraphic changes and no evidence of an unconformity as expected from a sea-level fall (Miller et al., 2008a). However, EOT-2 occurs in the upper part of the sequence associated with a shallowing-upward section, and estimates from Sites 1218 and 522 indicate a minor change in the $\delta^{18}\text{O}_{\text{seawater}}$ with EOT-2 (Katz et al., 2008). Thus, though we interpret EOT-2 as primarily a temperature drop, a sea-level component cannot be precluded.

The Oi-1 Increase

The greenhouse-to-icehouse transition culminated in the earliest Oligocene Oi-1 event (33.55 Ma), which is associated with an ~1‰ $\delta^{18}\text{O}$ increase in benthic foraminifera at SSQ and DSDP Site 522 and ODP Sites 744 and 1218. Global sea-level estimates derived from New Jersey backstripping (Figs. 1 and 2) show that a eustatic fall of 55 m occurred at ca. 33.5 Ma (Kominz and Pekar, 2001; Miller et al., 2005a; Kominz et al., 2008), correlated with the earliest Oligocene Oi-1 $\delta^{18}\text{O}$ increase. This decrease is equivalent to the growth of an Antarctic ice sheet of $25 \times 10^6 \text{ km}^3$, similar to the amount stored in East Antarctica today. The 55 m fall explains 0.55‰–0.66‰ of a global change in seawater $\delta^{18}\text{O}$ using the sea-level calibration of 0.10‰/10 m derived for the Oligocene (Pekar et al., 2002), the calibration of 0.11‰/10 m de-

rived for the Pleistocene (Fairbanks and Matthews, 1978), or the calibration of 0.12‰/10 m of Katz et al. (2008) for the Eocene-Oligocene. This indicates that the remaining ~0.34‰–0.45‰ of the increase at 33.55 Ma was due cooling of ~1.5–2 °C.

Uncertainties in the sea-level and/or $\delta^{18}\text{O}$ calibration potentially range from ~0.055‰ to 0.12‰/10 m (Miller et al., 1987; Pekar et al., 2002; Katz et al., 2008). The outside limits for any sea-level or $\delta^{18}\text{O}$ calibration are provided by the upper limit for freezing of ice sheets (~17‰) and the lower limit of the modern Antarctic ice sheet (~40‰), yielding a minimum calibration of 0.055‰/10 m (Miller et al., 1987) and a maximum of 0.12‰/10 m (Katz et al., 2008). The Pleistocene calibration is 0.11‰/10 m (Fairbanks and Matthews, 1978). Pekar et al. (2002) estimated an Oligocene calibration of 0.10 ± 0.02 ‰/10 m, similar to the value of 0.1‰/10 m obtained from climate models (DeConto and Pollard (2003). Katz et al. (2008) determined a calibration of 0.12‰/10 m for the Eocene-Oligocene transition. Thus, empirical calibration suggests that the average sea-level $\delta^{18}\text{O}$ calibration for the Cenozoic is 0.11 ± 0.1 ‰ (Fairbanks and Matthews, 1978; Pekar et al., 2002; Katz et al., 2008).

There are significant uncertainties in the sea-level estimate. Lowstand deposits are generally not represented in the coastal plain records used to construct the eustatic estimate (Miller et al., 2005a; Kominz et al., 2008). By conducting two-dimensional backstripping, Kominz and Pekar (2001) were able to estimate decreases during lowstand and thus reconstruct a nearly complete sea-level cycle for the earliest Oligocene sea-level lowering that correlates with Oi-1. Still, the upper limit of the eustatic fall could have been as high as 70 m. These estimates of a 55–70 m eustatic change do not take into account the isostatic response of the oceanic lithosphere to the change in weight of the overlying water (hydroisostasy). The actual water-volume change from transferring ocean water to glacial ice is ~33% higher than the eustatic change, assuming full isostatic compensation is attained. Hence, the 55–70 m eustatic change corresponds to a volume-equivalent change in the depth/thickness of ocean water of ~82–105 m (correcting for isostatic loading and assuming full compensation). Thus, the sea-level record only places moderate constraints on the ice-volume component of the $\delta^{18}\text{O}$ increase: anywhere from 55% to 100% of the 1.0‰ increase can be explained by a sea-level fall of 55–105 m.

Mg/Ca records place additional constraints on the $\delta^{18}\text{O}_{\text{seawater}}$ changes associated with the Oi-1 increase, though deep-sea Mg/Ca records may be overprinted by other effects (Lear et al., 2004; Katz et al., 2008). Both deep Atlantic DSDP Site 522 and deep Pacific ODP Site 1218 show no cooling or slight warming at Oi-1 time according to Mg/Ca paleothermometry (Lear et al., 2000, 2004). This observation led Tripathi et al. (2005) to attribute the bulk of the 1.5‰ $\delta^{18}\text{O}$ increase associated with the Eocene-Oligocene transition (i.e., both the EOT-1 and Oi-1 increase) to growth of ice sheets. However, the ~1.2 km drop in the CCD that occurred at the Eocene-Oligocene transition (van Andel, 1975; Coxall et al., 2005; Rea and Lyle, 2005) may have caused changes in carbonate ion activity that masked an estimated 2 °C cooling in

deep-sea Mg/Ca records (Lear et al., 2004). At SSQ, the Mg/Ca record supports 2 °C of cooling associated with the ~1.0‰ Oi-1 $\delta^{18}\text{O}$ increase, suggesting a 0.55‰ $\delta^{18}\text{O}_{\text{seawater}}$ change.

Thus, we conclude that the Eocene-Oligocene transition occurred in three major steps: (1) a precursor EOT-1 $\delta^{18}\text{O}$ increase at 33.80 Ma that was caused by a 2 °C deep-water cooling and a minor sea-level fall (~25 m) followed by a partial return to pre-event values; (2) a deep-water cooling and minor sea-level fall associated with the EOT-2 increase at 33.63 Ma; and (3) a deep-water cooling of 2 °C and a major sea-level fall of 80 ± 25 m at Oi-1 time (33.45 Ma).³ This implies the development of an Antarctic ice sheet that was as large or larger than that of present day. It does not require large Northern Hemisphere ice sheets. Edgar et al. (2007) similarly concluded that large Antarctic ice sheets predated large Northern Hemisphere ice sheets. However, it is likely given evidence from Arctic drilling (St. John, 2008) that at least Greenland-sized ice sheets existed during the Eocene and during times of peak glaciation such as Oi-1 (see following).

IMPLICATIONS OF A LARGE ANTARCTIC ICE SHEET

The cause of the earliest Oligocene oxygen isotope increase and the role of CO_2 (e.g., Pagani et al., 2005) versus changing ocean gateways (Kennett, 1977) have been hotly contested. The long-term $\delta^{18}\text{O}$ trend parallels atmospheric CO_2 estimates (Pagani et al., 1999, 2005), linking long-term (10^7 yr) global cooling to lower CO_2 (Fig. 1). Published modeling results (DeConto and Pollard, 2003) emphasized the importance of CO_2 on cryospheric development, where gateways played a role only in adjusting the critical $p\text{CO}_2$ threshold for Antarctic glaciation. These modeling studies suggest that a decrease in atmospheric CO_2 below a critical threshold (2.4–2.8 times pre-anthropogenic levels [280 ppm] for models with open and closed Drake Passage) in the earliest Oligocene could have triggered the growth of a large Antarctic ice sheet (Fig. 1), and sea-level estimates (Miller et al., 2005a; Kominz et al., 2008) that require a near modern-sized Antarctic ice sheet developed at ca. 33.55 Ma. DeConto et al. (2007) linked the growth of the Antarctic ice sheet to a CO_2 fall below this threshold, which is roughly consistent with the observed CO_2 estimates (Fig. 1). Though it is likely that there were small ($5\text{--}15 \times 10^6 \text{ km}^3$), ephemeral ice sheets during the greenhouse world of the Late Cretaceous to Eocene (Miller et al., 2005b), the earliest Oligocene saw the first continent-sized ice sheet since the Permian (ca. 280 Ma). Recent studies point toward much older Northern Hemisphere glaciation than was previously thought (e.g., middle Eocene; Moran et al., 2006; Eldrett et al., 2007), but the restriction of physical evidence for Northern Hemisphere glaciation (see summary in Wright and Miller, 1996) necessarily relegates it to a rela-

tively small role (e.g., similar to the modern-day 5–7 m sea-level equivalent stored in Greenland) until the last 2.6 m.y.

We suggest that the Antarctic ice sheet became a driver of, not just a response to, climate change beginning in the earliest Oligocene, causing increased meridional thermal gradients and a “spinning up” of the oceans (greater thermohaline and wind-driven circulation). Enhanced wind circulation due to katabatic winds resulted in increased wind shear/curl and hence increased circum-Antarctic upwelling. In addition, increased salt rejection due to ice-shelf and sea-ice freezing led to enhanced formation of Antarctic Bottom Water (AABW). Combined, these factors led to increased ventilation of the deep ocean, a decrease in residence time, and increased Southern Ocean productivity as discussed by Berger (2007). During the Eocene-Oligocene transition, the Antarctic ice sheet reached sufficient proportions to influence the ocean adjacent to the continent, resulting in increased sea ice and deep-water production. As first noted by Kennett and Shackleton (1976) and recently emphasized by models (DeConto et al., 2007), the formation of extensive sea ice likely played a role in changing deep-water circulation. There is considerable evidence that the Eocene-Oligocene transition was associated with the greatest change in deep-ocean circulation of the past 50 m.y. (Fig. 3). Thermohaline circulation dramatically increased in the early Oligocene (Fig. 3) with pulses of Southern Component Water (SCW; analogous to modern AABW; Kennett, 1977; Wright and Miller, 1996) and Northern Component Water (NCW; analogous to modern North Atlantic Deep Water) (Miller and Tucholke, 1983; Miller, 1992; Davies et al., 2001). An erosional event in the southern oceans caused hiatuses in numerous early Oligocene southern ocean cores and has been attributed to an increase in SCW (Kennett, 1977; Wright and Miller, 1996). The high supply of SCW is also indicated by the very high $\delta^{18}\text{O}$ values measured at high southern latitudes (Kennett and Stott, 1990), indicative of a “cold spigot” (Miller, 1992; Miller et al., 2008b). Seismic profiles and hiatus distributions document a coeval early Oligocene pulse of widespread erosion that cut unconformities associated with Horizon Au (Fig. 3; Mountain and Miller, 1992; Tucholke and Vogt, 1979) and reflector R4 (Miller and Tucholke, 1983) in the North Atlantic. This erosion correlates with high $\delta^{13}\text{C}$ values in the North Atlantic following the 33.55 Ma Oi-1 event (Fig. 3; Miller, 1992), supporting seismic and hiatus evidence that indicates a NCW source in the earliest Oligocene. Northern Hemisphere high latitudes cooled concomitantly with polar cooling in the Southern Hemisphere, making the North Atlantic a viable source region for deep water (albeit briefly) in the earliest Oligocene (see also Via and Thomas, 2006; Davies et al., 2001; Abelson et al., 2007).

A dramatic deepening of the CCD occurred across the Eocene-Oligocene transition (van Andel, 1975; Coxall et al., 2005; Rea and Lyle, 2005). Both the 33.80 (EOT-1) and 33.55 Ma (Oi1) $\delta^{18}\text{O}$ increases were associated with increases in carbonate content and mass accumulation rates in the deep Pacific (Coxall et al., 2005), and there appears to be a third, smaller increase associated with EOT-2 (Fig. 2). This indicates

³These estimates include errors in the eustatic estimate plus those due to water loading; e.g., the eustatic estimate for Oi-1 is 55–70 m, with water thickness increase of 55 m (i.e., no loading) to 105 m (i.e., full compensation).

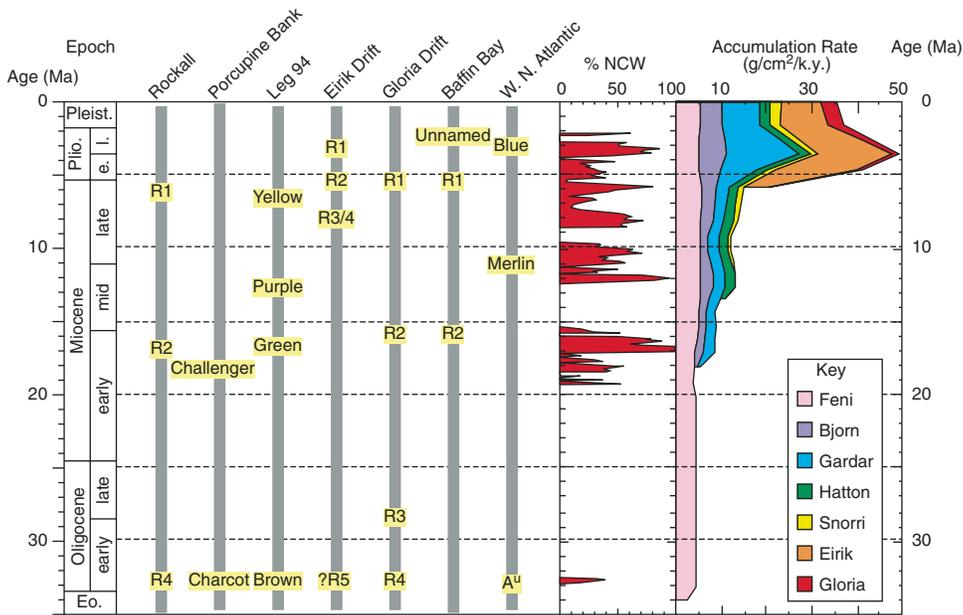


Figure 3. Age of North Atlantic seismic reflectors showing widespread erosional unconformities at ca. 34 Ma, ca. 20–16 Ma, ca. 13–10 Ma, and ca. 5 Ma; W.N.—Western North. Estimated flux of Northern Component Water (NCW) was derived using $\delta^{13}\text{C}$ records from the North Atlantic, Southern, and Pacific Oceans (Wright and Miller, 1996). Accumulation rates of northern North Atlantic drifts modified after Wright and Miller (1996). Note that increase in total drift accumulation corresponds to detection of NCW using $\delta^{13}\text{C}$ records.

that the CCD deepening occurred in two to three steps tightly coupled to the oxygen isotopic increases. We also note that although the drop in $\% \text{CaCO}_3$ at EOT-1 was as large as or larger than at Oi-1 time (Fig. 2), the carbonate accumulation rate data indicate that it was the latter event that was more significant. The deepening of the CCD can be attributed to one of the following: (1) an increase in deep-basin carbonate deposition at the expense of shallow carbonates, likely caused by sea-level fall (shelf-basin fractionation); (2) increased weathering of silicate rocks; (3) a global intensification of carbonate export to the deep sea, possibly caused by increased continental input to the oceans, or (4) a decrease of deep-ocean residence time.

Previous studies have attributed the CCD drop to a glacio-eustatic fall and a shift in shelf-basin fractionation (Coxall et al., 2005; Merico et al., 2008). We argue that shelf-basin fractionation cannot be invoked to explain the CCD drop because: (1) the sea-level fall associated with Oi-1 (33.55 Ma) was too small (80 ± 25 m) to account for the large deepening of the CCD (i.e., numerous Cenozoic sea-level events of this magnitude [Miller et al., 2005a] had little effect on the CCD); and (2) there was a CCD drop associated with EOT-1 despite that fact that there was only a small (~ 25 m) sea-level fall at that time. Though weathering may have increased in the earliest Oligocene (Robert and Kennett, 1997) and contributed to a global increase in carbonate production, global Oligocene sedimentation rates were low, and it is doubtful that input changes can explain the CCD drop. Rea and Lyle (2005) similarly argued convincingly that the CCD drop could not be entirely due to shelf-basin shift and suggested that a sudden increase in weathering and erosion rates would be unlikely to account for the change, thus implicating variations in deep-sea preservation. We argue that the links among deep-sea temperature, an increase in thermohaline circulation, and the

drop in the CCD are manifest as to its cause: increased thermohaline circulation and deep-basin ventilation caused a decrease in oceanic residence time and a decrease in deep-ocean acidity, allowing carbonate to be preserved at greater depth and produce a deeper CCD.

One reason that modeling studies (Merico et al., 2008) have favored shelf-basin fractionation for the CCD drop and ignored the importance of deep-sea ventilation is because they attempted to explain not only the CCD drop but also high global $\delta^{13}\text{C}$ values (e.g., Zachos et al., 1996). Carbon isotopic values began to increase at Oi-1 time, and high values lasted for ~ 1 m.y. (Zachos et al., 1996). Models showed that the CCD drop could be explained by an increase in organic carbon burial, an increase in silica over calcareous plankton deposition, or a shelf-basin fractionation, but only a transient event such as sea-level change could explain the transitory $\delta^{13}\text{C}$ peak (Merico et al., 2008). We note that the high $\delta^{13}\text{C}$ values must reflect a change in the input or burial of organic carbon relative to carbonate carbon and that such million-year-scale $\delta^{13}\text{C}$ cycles of $\sim 1\%$ typify the Oligocene to Miocene (Miller and Fairbanks, 1985). We argue that the high carbon isotopic values were not coupled directly to the CCD drop, and the $\delta^{13}\text{C}$ increase was due to a million-year-scale increase in organic carbon export relative to carbonate (as first noted by Zachos et al., 1996), and the CCD drop was due to increased deep-sea residence time.

Antarctic cooling coupled with tropical temperature stability (Pearson et al., 2007) implies a sharp increase in latitudinal thermal gradients (Shackleton and Kennett, 1975) that caused increased atmospheric circulation (Dupont-Nivet et al., 2007) and ocean upwelling (Berger, 2007). EOT-2 (33.63 Ma) corresponds to a major drop in nannofossil diversity in Tanzanian cores (Dunkley Jones et al., 2008), which is interpreted as a

response to surface-water eutrophication. The increased latitudinal and vertical thermal gradients caused a “spinning up” of the oceans and increased wind-driven upwelling and mesoscale eddy turbulence that contributed to greater nutrient recycling, which in turn stimulated a rapid diversification of diatoms (Falkowski et al., 2004a, 2004b; Katz et al. 2004; Finkel et al., 2005, 2007). As summarized by Katz et al. (2005), marine diatoms are favored over other eukaryotic phytoplankton by conditions with high turbulent mixing and thus are related to increased water-column stratification and latitudinal thermal gradients. Diatom species diversity remarkably mirrors the deep-sea $\delta^{18}\text{O}$ record across the Eocene-Oligocene transition (Fig. 1), as expected if deep-sea temperatures were also related to latitudinal thermal gradients and water-column stratification.

We note that diatom diversity increased not only at the Eocene-Oligocene transition but also in the middle Miocene and late Miocene–Pliocene (Fig. 1), when diversification included more bloom-forming diatoms (requiring high nutrient levels); these were also intervals of increased latitudinal thermal gradients and enhanced global circulation (Wright and Miller, 1996). Our comparisons (Fig. 1) indirectly link silica productivity to these intervals of global cooling and increased latitudinal thermal gradients.

As hypothesized by DeConto and Pollard (2003), atmospheric CO_2 , which declined below a critical threshold in the earliest Oligocene, may have propelled the climate system into a new state, one with a large Antarctic ice sheet. We suggest the continent-sized ice sheet provided positive feedback that was felt throughout the climate system, probably through the profound effects of sea ice on ocean circulation and chemistry. Though there are large uncertainties in atmospheric CO_2 estimates (Fig. 1), it appears that the broad-scale trends in Cenozoic CO_2 and the long-term (10^7 yr scale) sea-level record (Fig. 1) track each other. The first-order trends of decreasing atmospheric CO_2 and falling sea level appear to have been coupled since 50 Ma, strongly implying a link through changes in ocean crust production rates (Larson, 1991; Miller et al., 2005a). The global long-term cooling of $\sim 12^\circ\text{C}$ associated with the lowering of CO_2 can only explain 12 m of steric sea-level fall, and therefore the bulk of the long-term fall of ~ 100 m (Kominz et al., 2008) must have been due to: (1) long-term ice growth and isostatic compensation (~ 50 m); and (2) lower ocean crust production rates (~ 40 m). Higher estimates of long-term fall (e.g., Müller et al., 2008) would require greater reduction in ocean crust production rates. The cause of the drop in CO_2 is less clear, and it may be attributed to lower degassing due to lower ocean crust production rates and/or possibly to higher continental weathering rates. The correlation between long-term CO_2 and long-term sea level (Fig. 1) suggests that they are linked through ocean crust production rates.

This evaluation of the Eocene-Oligocene transition provides perspective on the evolution of Cenozoic climate. Oxygen isotope shifts that occur through the Eocene-Oligocene transition (Fig. 1), in the middle Miocene (Fig. 1), and in the mid-Pliocene overshoot estimates of CO_2 and sea-level change, suggesting am-

plification within the climate system. This implies that CO_2 was not the only driver of Cenozoic climate change, and it leaves open the possibility that CO_2 changes were augmented by a positive feedback mechanism for climate change (e.g., large Antarctic ice sheets, increasing latitudinal thermal gradients, and ocean circulation, resulting in additional CO_2 drawdown and cooling). These events mark major changes in the climate state that were amplified by processes other than changes in CO_2 , as first suggested in a prescient paper by Berger (1982). In the case of these three events, it appears that the development of continental-scale ice sheets was an important factor in the amplification of the climate system: (1) in the Eocene-Oligocene transition, the development of a continental-scale ice sheet on Antarctica resulted in major feedbacks through thermohaline circulation and ocean upwelling; (2) in the middle Miocene, the development of a permanent (i.e., dry-based) ice sheet in Antarctica further increased latitudinal thermal gradients (Wright and Miller, 1996); and (3) in the Pliocene, the growth of large Northern Hemisphere ice sheets exerted another amplification, perhaps through a reduction in NCW.

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