Cooling and ice growth across the Eocene-Oligocene transition

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ABSTRACT

The Eocene-Oligocene (E-O) climate transition (ca. 34 Ma) marks a period of Antarctic ice growth and a major step from early Cenozoic greenhouse conditions toward today's glaciated climate state. The transition is represented by an increase in deep-sea benthic foraminiferal oxygen isotope ($\delta^{18}O$) values occurring in two main steps that reflect the temperature and $\delta^{18}O$ of seawater. Existing benthic Mg/Ca paleotemperature records do not display a cooling across the transition, possibly reflecting a saturation state effect on benthic foraminiferal Mg/Ca ratios at deep-water sites. Here we present data from exceptionally well preserved foraminifera deposited well above the calcite compensation depth that provide the first proxy evidence for an ~2.5 °C ocean cooling associated with the ice growth. This permits interpretation of E-O $\delta^{18}O$ records without invoking Northern Hemisphere continental-scale ice.

Keywords: Eocene, Oligocene, climate, ice sheets, temperature, Cenozoic.

INTRODUCTION

The timing and magnitude of Earth's climatic changes through the Cenozoic are well documented by deep-sea benthic foraminiferal oxygen isotope ($\delta^{18}O$) records (e.g., Zachos et al., 2001). These records reflect variations in continental ice volume because the hydrologic cycle concentrates ^{16}O in ice sheets, leaving the oceans more enriched in ^{18}O as glaciations advance. However, benthic $\delta^{18}O$ records also contain a temperature component associated with isotopic fractionation during calcification. Therefore, without an independent proxy for either ice volume or temperature, these two components of the benthic $\delta^{18}O$ records cannot be quantified. Benthic foraminiferal Mg/Ca paleothermometry is a salinity-independent temperature proxy that can be used to deconvolve the ice volume signal from benthic foraminiferal $\delta^{18}O$ records (e.g., Lear et al., 2000).

The Eocene-Oligocene (E-O) climate transition marks the establishment of the Antarctic ice sheet ca. 33–34 Ma, and has been associated with cooling of low-, middle-, and high-latitude continents (see recent review in Coxall and Pearson, 2007). From an ocean perspective, benthic foraminiferal $\delta^{18}O$ records display a two-stepped ~1.5‰ increase (Coxall et al., 2005). Global climate models that simulate the establishment of the Antarctic ice sheet under scenarios of decreasing atmospheric pCO_2 predict an increase in mean seawater $\delta^{18}O$ of ~0.5‰ (DeConto and Pollard, 2003). The remainder of the ~1.5‰ shift in the deep-sea foraminiferal records has been attributed to oceanic cooling and/or accumulation of continental ice outside of Antarctica (Coxall et al., 2005).

Previous attempts to produce $\delta^{18}O$-independent paleotemperature reconstructions using Mg/Ca paleothermometry have failed to find evidence for ocean cooling associated with the E-O climate transition (Lear et al., 2000, 2004; Billups and Schrag, 2003), raising the possibility that there were substantial Northern Hemisphere ice sheets in the earliest Oligocene (Coxall et al., 2005). There are, however, a number of problems with the idea of widespread Northern Hemisphere glaciation at this time. First, current coupled general circulation model–ice sheet models are only able to produce major bipolar glaciation events during the early Cenozoic when using pCO_2 levels near or below modern-day values (DeConto and Pollard, 2007), yet proxy based pCO_2 estimates suggest that atmospheric pCO_2 concentrations were at least twice modern levels during late Eocene to early Oligocene time (Pagani et al., 2005). Second, although recent sedimentological evidence supports the existence of Eocene–Oligocene ice on Greenland, the relative scarcity of glacial debris in North Atlantic sediments supports the presence of small ephemeral ice caps and valley glaciers rather than a more substantial continental-scale ice sheet (Eldrett et al., 2007). There is, at present, no compelling sedimentological evidence to support the existence of substantial Northern Hemisphere continental-scale ice during the E-O climate transition.

An alternative explanation for the mismatch between the data and the model is that there is a problem with current E-O Mg/Ca paleotemperature records because of the >1 km deepening of the calcite compensation depth (CCD) (Rea and Lyle, 2005) that occurred contemporaneously with the $\delta^{18}O$ shift across the E-O transition (Coxall et al., 2005). This deepening of the CCD reflects a rapid change in the carbonate saturation state of the deep ocean, which might have affected the partitioning of Mg into benthic foraminiferal calcite (Lear et al., 2004; Elderfield et al., 2006). Here we test this hypothesis by generating seawater temperature records (Mg/Ca and Sr/Ca paleothermometry) from a site situated well above the paleo-CCD.

MATERIALS AND METHODS

The Tanzanian Drilling Project (TDP) has recovered three overlapping cores across the E-O transition (TDP11, TDP12, and TDP17) (Pearson et al., 2008). TDP cores were logged in the field and a half-round sample covering a 10 cm interval was taken from every section, representing a resolution of ~1 sample/m. Further details demonstrating the completeness of the sections as deduced from biostratigraphy and sedimentary features are available in Pearson et al. (2008). Benthic foraminiferal assemblages suggest a paleowater depth of a few hundred meters, and planktonic assemblages indicate normal marine conditions. The TDP sites were situated ~50 km from the paleoshoreline (Nicholas et al., 2006).

The TDP E-O boundary sections contain the benthic $\delta^{18}O$ isotope shift and biostratigraphic horizons, including the extinction horizon of *Hantkenina* spp., which marks the E-O boundary worldwide (Pearson et al., 2008). An age model based on planktonic foraminifera and nannofossil biostratigraphy linked to the Cande and Kent (1995) time scale has been produced (Pearson et al., 2008). The Tanzanian cores have four distinct advantages over typical deep-sea sediments in producing geochemical proxy records of the E-O climate transition. (1) Unlike many deep open ocean sites, the hemipelagic clay lithology has resulted in extremely well preserved (glassy) foraminiferal calcite (Pearson et al., 2007). (2) The paleowater depth (lower shelf to slope facies, a few hundred meters water depth; Nicholas et al., 2006) is well above the lysocline, as evidenced by abundant aragonitic benthic foraminifera showing no visible signs of dissolution. Therefore dissolution and/or carbonate saturation effects on the foraminiferal trace metal geochemistry can be considered negligible. (3) This paleowater depth is also shallow enough that a eustatic sea-level lowering may have an impact on bottom-water temperatures. Therefore, changes in the planktonic-benthic temperature gradients potentially provide an additional means of delimiting the timing of ice sheet growth. (4) Tropical sea-surface temperature records that are sited far from the influence of high-latitude ice sheets can be used to distinguish between global cooling versus regional cooling resulting from ice sheet growth.

Prior to trace metal analysis, foraminifera were cleaned using a rigorous protocol adapted from Boyle and Keigwin (1985). The procedure included a clay removal step, a manual picking out of contaminant phases using a binocular microscope (Barker et al., 2003), two oxidation steps, two reductive steps, and four acid leaches. Foraminiferal Mg/Ca and Sr/Ca ratios were analyzed by high-resolution–inductively coupled plasma–mass spectrometry (long-term precision <2% relative standard deviation) Benthic foraminifera were analyzed at Rutgers University, and planktonic foraminifera were analyzed at Bristol University. All trace metal records were analyzed on samples from TDP12; data are available in the GSA Data Repository[1]. *Cibicidoides* specimens were assigned to species A, B, or C, and analyzed as monospecific samples. Photographs of these three species were taken using a scanning electron microscope and are also available (see the GSA Data Repository).

Sea-surface temperatures were calculated from Mg/Ca ratios of the planktonic foraminifer *Turborotalia ampliapertura* (a near-surface dweller) using the generic Mg/Ca-temperature equation of Anand et al. (2003), assuming a seawater Mg/Ca ratio of 4.3 mol/mol (Wilkinson and Algeo, 1989). The calibration of Anand et al. (2003) is based on several modern planktonic species from evolutionary groups that diverged prior to the E-O transition. We note that *T. ampliapertura* is within the phylogenetic bracket of the modern groups, suggesting that although the species is extinct, the calibration is appropriate. Combined $\delta^{18}O$-$\delta^{13}C$ data demonstrate that *T. ampliapertura* is a near-surface species with $\delta^{18}O$ isotopic temperatures heavier by an equivalent of 2 °C relative to forms living in the uppermost surface layers (B.S. Wade, 2007, personal commun.). Therefore we add 2 °C to our calculated near-surface temperatures to obtain sea-surface temperatures. It is reassuring that these estimates are similar to sea-surface temperatures estimated using the TEX_{86} paleotemperature proxy, which is derived independently from organic biomarkers (Pearson et al., 2007; Fig. 1). Near-surface seawater $\delta^{18}O$ ($\delta^{18}O_{sw}$) was estimated using the near-surface temperatures and coincident $\delta^{18}O$ of *T. ampliapertura* following Erez and Luz (1983).

Bottom-water temperatures were calculated from Sr/Ca ratios of the aragonitic benthic foraminifer *Epistomina* sp., which is sometimes referred to as *Hoeglundina elegans* (van Morkhoven et al., 1986). A recent Sr/Ca-temperature calibration for *H. elegans* (Rosenthal et al., 2006) yields unrealistically large temperature variations in comparison

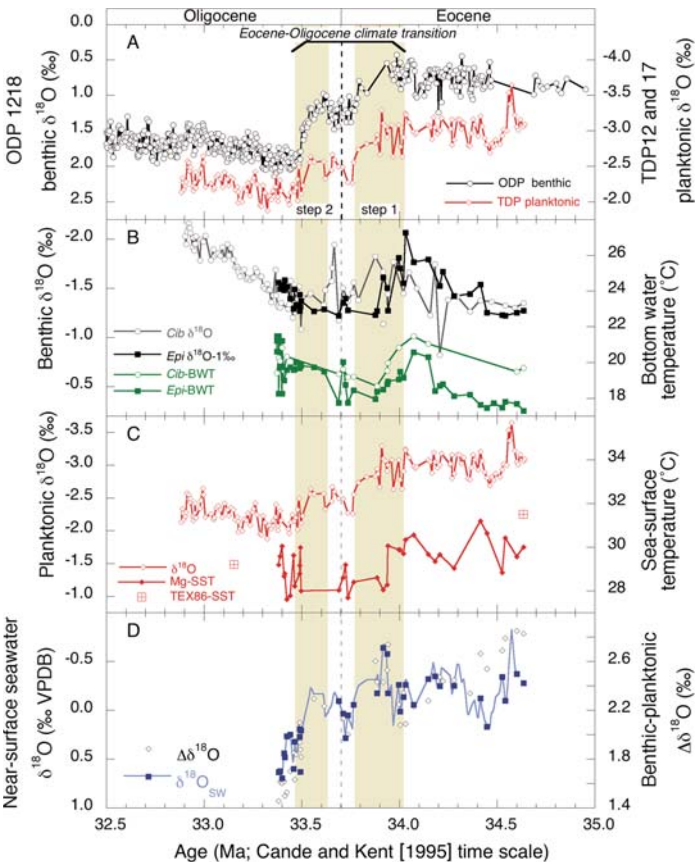

Figure 1. Multiproxy geochemical records of ocean and climate change across the Eocene-Oligocene (E-O) transition. A: Oxygen isotope records. Black circles—benthic foraminiferal $\delta^{18}O$ data from Ocean Drilling Program (ODP) Site 1218 (Coxall et al., 2005); red diamonds—planktonic foraminiferal $\delta^{18}O$ data from Tanzanian Drilling Project (TDP) Sites 12 and 17 (Pearson et al., 2008). B: Benthic foraminiferal $\delta^{18}O$ data and bottom-water temperatures from TDP Sites 12 and 17. Open gray circles—multispecific *Cibicidoides* spp. $\delta^{18}O$; solid black squares—monospecific *Epistomina* sp. $\delta^{18}O$ adjusted by −1‰; open green circles—*Cibicidoides* sp. Mg/Ca temperatures; solid green squares—*Epistomina* sp. Sr/Ca temperatures. C: Sea-surface temperature (SST) records from TDP Sites 11, 12, and 17. Closed red diamonds—calculated from *Turborotalia ampliapertura* Mg/Ca from TDP Site 12; hachured red squares—derived from TEX_{86} measurements from TDP Site 11 (Pearson et al., 2007); open red diamonds—TDP composite planktonic foraminiferal $\delta^{18}O$ record as in A (Pearson et al., 2008). D: Calculated near-surface $\delta^{18}O_{sw}$ offshore Tanzania across E-O transition. Blue line—calculated by interpolating both planktonic $\delta^{18}O$ and Mg/Ca-temperature records (C); solid blue squares—calculated using planktonic Mg/Ca temperatures and interpolating planktonic $\delta^{18}O$ record; open black diamonds—benthic-planktonic $\delta^{18}O$ gradient, in part reflecting water depth, calculated from *Epistomina* sp. $\delta^{18}O$ and interpolated *T. ampliapertura* $\delta^{18}O$. Yellow shaded bars highlight the two "steps" of the oxygen isotope shift (Coxall et al., 2005), and vertical dashed black line marks E-O boundary as defined by extinction of the Family Hantkeninidae. VPDB—Vienna Peedee belemnite.

[1]GSA Data Repository item 2008061, aragonite Sr/Ca temperature calibration, scanning electron microscope images, foraminiferal trace metal data and calculated temperatures, is available online at www.geosociety.org/pubs/ft2008.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

to the *Epistomina* $\delta^{18}O$ and planktonic Mg/Ca-temperature records. This is possibly a result of a bias in the Rosenthal et al. (2006) calibration toward core-top samples from Little Bahama Bank that may be altered diagenetically (Marchitto et al., 2007; Lear et al., 2002). Subsequent results from other Atlantic sites are consistent with the Pacific calibration and not with results from Little Bahama Bank (Y. Rosenthal, 2007, personal commun.). Therefore, we use a calibration based solely on *H. elegans* collected from oversaturated waters from the Indonesian Seaway (Rosenthal et al., 2006; see the GSA Data Repository) that yields temperature variations that are consistent with the benthic $\delta^{18}O$ and the planktonic Mg/Ca records. To calculate absolute Sr/Ca paleotemperatures, we assume that seawater Sr/Ca was 17% lower than modern values during the E-O transition (Lear et al., 2003). Bottom-water temperatures were also calculated from Mg/Ca ratios of three monospecific records of the calcitic benthic foraminifer genus *Cibicidoides*, using the exponential calibration of Lear et al. (2002), and assuming seawater Mg/Ca = 4.3 mol/mol (Wilkinson and Algeo, 1989).

While the absolute temperatures calculated from trace metal records are dependent on vital effects and estimates for seawater Mg/Ca and Sr/Ca, estimates of relative temperature changes across short-lived events such as the E-O transition are independent of these assumptions. However, we note that the calculated relative temperature changes are sensitive to the choice of calibration, which is an additional source of uncertainty particularly for the benthic Mg/Ca temperatures. For example, a linear benthic Mg/Ca-temperature calibration based on a modern Florida Straits species produces unrealistically large temperature variations for our down-core records (Marchitto et al., 2007). Nevertheless, the absolute temperatures that we have calculated from the different proxies (planktonic Mg/Ca versus TEX_{86}, and calcite benthic Mg/Ca versus aragonite benthic Sr/Ca) are in relatively good agreement (Fig. 1).

E-O Stable Isotope Stratigraphy

The TDP planktonic $\delta^{18}O$ record, which represents an overlapping stratigraphy from cores TDP12 and TDP17 (Pearson et al., 2008), shows a pattern of $\delta^{18}O$ increase across the E-O transition similar to that of deep-sea Pacific records (Coxall et al., 2005), demonstrating that the TDP planktonic foraminifera record a global climate signal (Fig. 1A). Planktonic foraminiferal $\delta^{18}O$ increases by ~0.7‰ across the first step, and ~0.4‰ across the second step. The monospecific *Epistomina* sp. benthic $\delta^{18}O$ record displays an increase across the first step of the same order of magnitude as the planktonic $\delta^{18}O$ record. However, the *Epistomina* sp. $\delta^{18}O$ record does not display an increase across the second step of the climate transition, presumably because this record also reflects bathymetric changes of the shelf-slope environment caused by glacio-eustatic sea-level change (Fig. 1B). The multispecific *Cibicidoides* spp. $\delta^{18}O$ record is relatively noisy across the E-O climate transition, possibly a result of varying vital effects between species, which were combined for isotopic analysis (Fig. 1B).

TEMPERATURE, ICE VOLUME, AND SEA-LEVEL CHANGE

Prior to the E-O climate transition, in the interval between ca. 34.4 and 34.1 Ma, the planktonic-benthic foraminiferal temperature gradients decrease as bottom waters warmed (Figs. 1B–1D). This decrease in near-surface to bottom-water temperature contrasts cannot be attributed to glacio-eustatic lowering of sea level because calculated seawater $\delta^{18}O$ remains more or less constant through this interval (Fig.1D). While we cannot rule out a regional tectonic effect on local sea level, we suggest it most likely that the warming of the shelf-slope bottom waters reflects a deepening of the thermocline, perhaps associated with reduced stratification.

The initial ~0.7‰ planktonic $\delta^{18}O$ step of the E-O climate transition coincides with a decrease in the planktonic Mg/Ca values, the aragonitic benthic Sr/Ca values, and the calcitic benthic Mg/Ca values, equivalent to an ~2.5 °C cooling of tropical near-surface and bottom waters (Figs. 1B, 1C). This implies that the majority (~0.5‰) of the $\delta^{18}O$ in step 1 represents cooling, with only the remaining ~0.2‰ available to accommodate an increase in continental ice volume (Fig. 1D). Consistent with this is the observation that the magnitude of the initial isotope step in the tropical near-surface waters is similar to that recorded by deep ocean foraminifera (and hence high-latitude surface waters) (Fig. 1A).

Unlike the initial planktonic foraminiferal isotope increase, the second step in $\delta^{18}O$ is not associated with a decrease in near-surface temperatures, suggesting instead that this shift reflects an increase in $\delta^{18}O_{sw}$ of ~0.4‰ (Figs. 1C, 1D). We therefore estimate the total increase in $\delta^{18}O_{sw}$ across the climate transition into the early Oligocene glacial maximum as ~0.6‰. Translating this value into an ice-volume equivalent requires estimation of the isotopic composition of ancient ice sheets (ice sheets in warmer climates could have had heavier $\delta^{18}O$ than today), but most likely represents an increase in ice volume approximately equivalent to the modern-day Antarctic ice sheet (Table 1). The associated ~70 m of apparent sea-level fall estimated from sequence stratigraphy (Pekar et al., 2002) is consistent with the growth of an ice sheet of this size.

The suggestion that the second $\delta^{18}O$ shift is dominated by a large ice volume component is supported by the associated decrease in the Tanzanian benthic-planktonic temperature gradient, with benthic temperatures becoming warmer as shelf-slope waters shoaled in response to the eustatic sea-level fall (Figs. 1B, 1D). Further support for this interpretation is provided by the observation that the magnitude of the second $\delta^{18}O$ step is greater in the deep-sea records than the tropical surface record (Pearson et al., 2008; Fig. 1A). As the Antarctic ice sheet approached the edge of the continent, a regional cooling effect might be expected in the Southern Ocean. In addition, it is possible that significant sea ice cover in the Southern Ocean initiated at that time, which may also have cooled Southern Ocean waters (DeConto et al., 2007). This cooling may have been transmitted to the deep ocean by water masses sourced from high southern latitudes. However, we also note the possibility that a component of the cooling observed in the deep-sea records may represent a change in the location of the source of deep-water masses.

CONCLUSIONS

If our interpretation of the Tanzanian record is correct, it seems likely that the increase in deep ocean carbonate saturation state associated with the E-O climate transition (Coxall et al., 2005) masked the effects of cooling in the Ocean Drilling Program Site 1218 Mg/Ca-temperature record (Lear et al., 2004). The latter record is based on sediments recovered very close to the critical depth of CCD shift (~3.6 km paleo-water depth). Although the E-O CCD deepening represents an unusually dramatic change in deep ocean saturation state, this work highlights the importance for choosing appropriate material for benthic foraminiferal Mg/Ca paleothermometry when coeval changes in seawater saturation state are suspected (Lear et al., 2004, Elderfield et al., 2006).

TABLE 1. CALCULATED ICE VOLUMES†

Assumed average composition of glacial ice	Calculated ice volume (km^3)	Total modern ice volume (%)	Modern Antarctic ice volume (%)
–50‰	1.69×10^7	63	70
–45‰	1.88×10^7	71	78
–40‰	2.12×10^7	80	88
–35‰	2.42×10^7	100	110
–30‰	2.84×10^7	107	118

†Volumes required to produce a 0.6‰ increase in seawater $\delta^{18}O$, assuming an ice-free world has a mean oceanic isotopic composition of –1.2‰ and a volume of 1.39×10^9 km^3 (Shackleton and Kennett, 1975). Mean composition of modern Antarctic ice is ~–50‰ (Shackleton and Kennett, 1975).

Recent studies agree that the establishment of the Antarctic ice sheet was paced by changes in Earth's orbit (Coxall et al., 2005; DeConto and Pollard, 2003), with the underlying climatic preconditioning factor variously ascribed to high-latitude cooling (Kennett, 1977), changes in moisture transport onto Antarctica (Lear et al., 2000; Bartek et al., 1992), or global cooling, either gradual or sudden (DeConto and Pollard, 2003; Zachos and Kump, 2005). Our results indicate a tropical surface-water temperature decrease (~2.5 °C) and support a global-cooling mechanism. The total increase in continental ice budget across the E-O transition is approximately equivalent to the modern-day Antarctic ice sheet, and can be accounted for without invoking continental-scale Northern Hemisphere glaciation during a period when atmospheric pCO_2 levels are generally believed to have been substantially higher than today (Pagani et al., 2005).

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