Early Oligocene glaciation and productivity in the eastern equatorial Pacific: Insights into global carbon cycling

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[1] The onset of sustained Antarctic glaciation across the Eocene-Oligocene transition (EOT) marks a pivotal change in Earth's climate, but our understanding of this event, particularly the role of the carbon cycle, is limited. To help address this gap we present the following paleoceanographic proxy records from Ocean Drilling Program Site 1218 in the eastern equatorial Pacific (EEP): (1) stable isotope (δ^{18} O and δ^{13} C) records generated in epifaunal benthic foraminifera (Cibicidoides spp.) to improve (double the resolution) the previously published records; (2) δ^{18} O and δ^{13} C records measured on Oridorsalis umbonatus, a shallow infaunal species; and (3) a record of benthic foraminifera accumulation rate (BFAR). Our new isotope data sets confirm the existence at Site 1218 of a two-step δ^{18} O increase. They also lend support to the hypothesized existence of a late Eocene transient δ^{18} O increase and early Oligocene Oi-1a and Oi-1b glacial maxima. Our record of BFAR indicates a transient (~500 kyr) twofold to threefold peak relative to baseline Oligocene values associated with the onset of Antarctic glaciation that we attribute to enhanced biological export production in the EEP. This takes the same general form as the history of opal accumulation in the Southern Ocean, suggesting strong high-to-low-latitude oceanic coupling. These findings appear to lend support to the idea that the EOT δ^{13} C excursion is traceable to increased organic carbon (Corg) burial. Paradoxically, early Oligocene sediments in the EEP are extremely Corg-poor, and proxy records of atmospheric pCO2 indicate a transient increase associated with the EOT.

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1. Introduction

[2] The onset of sustained Antarctic glaciation close to the Eocene-Oligocene (E-O) boundary represents a fundamental step in the evolution of our modern glacial climate state. The principal evidence for glacial expansion at this time comes from a large (1.0%-1.5%) increase in deep-sea oxygen stable isotopes (δ^{18} O) coincident with sedimentary evidence for extensive ice sheets reaching sea level across much of Antarctica [*Shackleton and Kennett*, 1975; *Kennett and Shackleton*, 1976; *Miller et al.*, 1987, 2009; *Ehrmann and Mackensen*, 1992; *Sagnotti et al.*, 1998; *Zachos et al.*, 1999, 2001; *Ivany et al.*, 2006; *Coxall and Pearson*, 2007; *Expedition 318 Scientists*, 2010].

[3] The cause of the Eocene-Oligocene climate transition (EOT) is much debated, with explanations falling into two broad categories: (1) those invoking changes in ocean/ atmosphere heat transport associated with tectonic breaching of ocean gateways and (2) those involving weakening of the

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Eocene greenhouse because of declining atmospheric carbon dioxide (pCO₂), combined with orbital configurations that favored ice sheet growth [e.g., *Kennett and Shackleton*, 1976; *DeConto and Pollard*, 2003; *Huber and Nof*, 2006; *Sijp and England*, 2004; *Livermore et al.*, 2005; *Zachos and Kump*, 2005; *Pagani et al.*, 2005; *Coxall et al.*, 2005; *Huber and Nof*, 2006; *Pearson et al.*, 2009; *Cramer et al.*, 2009]. Biotic and climatic responses to E-O climate changes were severe and widespread and include accelerated evolutionary turnover on land and in the oceans, 4°C–5°C cooling at high and low latitudes, aridification of continents and changes in weathering [*Ehrmann and Mackensen*, 1992; *Zachos et al.*, 1999, 2001; *Ravizza and Peucker-Ehrenbrink*, 2003; *Dalai et al.*, 2006; *Dupont-Nivet et al.*, 2007; *Lear et al.*, 2008; *Liu et al.*, 2009].

[4] Evidence that Earth's carbon system changed dramatically during the EOT comes from a variety of geochemical and sedimentary sources and includes: (1) an approximately 1‰ positive excursion in benthic δ^{13} C lasting ~1 million years (Myr), (2) an up to 1.3 km deepening of the calcite compensation depth (CCD), and (3) changes in biogenic marine sediment accumulation [*Diester-Haass and Zahn*, 1996; *Salamy and Zachos*, 1999, *Vanden Berg and Jarrard*, 2004; *Coxall et al.*, 2005; *Ravizza and Paquay*, 2008]. A decrease in atmospheric pCO₂ is also suggested [*Pearson et al.*, 2009], but the timing and structure of this

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change is still poorly constrained. These observations imply that the carbon system played a crucial role in the abrupt switch in climate mode but the mechanisms linking these changes are not well understood.

[5] Current hypotheses linking the carbon cycle and climate system responses for the EOT invoke feedbacks among atmospheric pCO_2 and (1) ice sheet coverage and silicate weathering rates [Zachos et al., 1996; Ravizza and Peucker-Ehrenbrink, 2003; Zachos and Kump, 2005; Dalai et al., 2006], (2) increased marine organic carbon burial/cycling [Diester-Haass and Zahn, 1996; Salamy and Zachos, 1999; Dalai et al., 2006; Ravizza and Paquay, 2008; Miller et al., 2009], (3) a global switch in the ecology of plankton to favor siliceous over calcareous organisms [Nilsen et al., 2003; Falkowski et al., 2004; Coxall et al., 2005; Zachos and Kump, 2005; Merico et al., 2008], (4) a shift in global carbonate sedimentation from the shelf to the deep ocean [Berger and Winterer, 1974; Opdyke and Wilkinson, 1989; Coxall et al., 2005; Merico et al., 2008], (5) changing riverine chemical inputs [Salamy and Zachos, 1999; Rea and Lyle, 2005; Merico et al., 2008], (6) an increase in the size or mean isotopic composition of the terrestrial biosphere [Salamy and Zachos, 1999], (7) increases in the rain ratio of inorganic:organic carbon from the surface to the deep ocean [Griffith et al., 2010], and (8) reduced ocean acidity because of basin ventilation related to increased ocean overturning [Miller et al., 2009]. The diversity of these hypotheses, combined with the variety of geochemical modeling outcomes serves to highlight large gaps in our understanding of the contribution of the various carbon system components, and the size and duration of external forces and feedbacks, such as weathering, circulation change and sea level fall.

[6] One gap in knowledge concerns the extent and timing of E-O changes in ocean productivity, export production and organic carbon burial. This is an important weakness to address because of the role played by marine biological carbon pumping in modulating atmospheric CO₂, and therefore climate. Changes in phytoplankton communities, increased sedimentary opal and increased numbers of benthic organisms seen in Southern ocean records, suggest higher primary production in the high southern latitudes from the middle Eocene and a further sharp increase near the E-O boundary [e.g., *Wei and Wise*, 1990; *Ehrmann and Mackensen*, 1992; *Salamy and Zachos*, 1999; *Diester-Haass and Zahn*, 1996; *Schumacher and Lazarus*, 2004; *Anderson and Delaney*, 2005].

[7] Outside the Southern Ocean, productivity increases across the EOT have been reported in the South Atlantic [Hartl et al., 1995], the equatorial Atlantic [Diester-Haass and Zachos, 2003], NW Australia [Diester-Haass and Zahn, 2001] and a number of sites close to continental margins [Ravizza and Paquay, 2008]. The majority of these studies, however, are of low resolution and lack sufficient age control to assess the relationship with the intensification of Antarctic glaciation. Moreover, the role of the eastern equatorial Pacific (EEP), which today supports a significant proportion (18%–56%) of global new production and is responsible for about two thirds of ocean-to-atmosphere CO2 release [Chavez and Toggweiler, 1995; Takahashi et al., 2002, 2009] is virtually unknown. Existing data sets from the EEP over the EOT interval are sparse and point to different interpretations of change in productivity. Patterns of bulk sediment accumulation have been interpreted as signaling no relative increase in EEP export production and burial [Schumacher and Lazarus, 2004, Vanden Berg and Jarrard, 2004; Griffith et al., 2010], yet a spike in diatom abundance close to the E-O boundary recorded in several equatorial Pacific sites [Fenner, 1984; Baldauf, 1992; Barron et al., 2004] suggests at least a short-lived increase in surface production, involving increased silica supply, associated with widespread Antarctic glaciation. Chronologic and temporal resolution of these existing diatom records, which are derived from discontinuous rotary-drilled deep-sea cores, however, is insufficient for making direct comparisons to high-resolution proxy records of glaciation.

[8] Scientific ocean drilling (ODP Leg 199 [Lyle et al., 2002]; IODP Expedition 320 [Pälike et al., 2009]) has greatly improved spatial and stratigraphic coverage of EOT sections in the EEP. Our study capitalizes on the high-quality sequence recovered at ODP Site 1218 (Leg 199) to improve on the epifaunal benthic foraminifera stable isotope record published by *Coxall et al.* [2005] and add infaunal stable isotope and BFAR records to gain further insight into carbon cycling across the EOT.

[9] Here we report three new data sets based on benthic foraminifera for the EOT from ODP Site 1218: (1) epifaunal benthic stable isotope (δ^{18} O and δ^{13} C) records based on *Cibicidoides* spp. that double the resolution of those published by *Coxall et al.* [2005], (2) accompanying δ^{18} O and δ^{13} C records for the shallow infaunal species *O. umbonatus*, and (3) a record of benthic foraminifera accumulation rate (BFAR) in two size fractions. Our principal aim is to test the hypothesis that the onset of widespread Antarctic glacial activity in the earliest Oligocene was associated with a change in primary productivity and export production in the EEP.

2. Oceanographic and Geological Setting

[10] ODP Leg 199 drilled a series of holes along a latitudinal transect in the equatorial Pacific to obtain climate records through the Paleogene [*Lyle et al.*, 2002]. Site 1218 is situated in the EEP (Figure 1) on a basement swell ~3°N of the Clipperton Fracture Zone [*Shipboard Scientific Party*, 2002]. Sediments of EOT age are not deeply buried at this site (~215 m below seafloor) because northeastern motion of the Pacific plate carried the site away to abyssal depths where red clay accumulation was slow (~0.5 cm kyr⁻¹). The recovered cores, therefore, contain mostly pelagic oozes, typically radiolarites in the Eocene, switching to nannofossil ooze in the early Oligocene. A fixed hot spot model indicates that the site remained within the equatorial high-productivity zone (within 2° of the paleo-equator) during the E-O interval [*Moore et al.*, 2004; *Parés and Moore*, 2005].

[11] Site 1218 has proved to be an excellent reference for E-O climate studies for a variety of reasons: (1) shipboard physical property measurements and the results of downhole logging reveal excellent stratigraphic coverage with no obvious hiatuses through the EOT in the three holes drilled; (2) Site 1218 is critically positioned at a paleo-depth of ~3800 m and thus captures the ~1300 m CCD deepening (from ~3300 to ~4500 m) associated with the EOT [*Lyle et al.*, 2002; *Rea and Lyle*, 2005; *Coxall et al.*, 2005]; (3) biostratigraphy, magnetostratigraphy, and cyclostratigraphy provide excellent stratigraphic control that have



Figure 1. Modern location of ODP Site 1218 and other key sites discussed in this paper. ODP Site 1218 is at a depth of 4826 m. TDP is the Tanzania Drilling Project.

permitted the construction of an astronomically tuned age framework [*Coxall et al.*, 2005; *Pälike et al.*, 2006]; and (4) from ~35.2 million years ago onward, sufficient numbers of deep-sea benthic foraminifera are present to allow generation of deep-sea paleoclimate proxy records.

3. Methods

3.1. Samples and Isotopic Analysis

[12] Twenty cubic centimeter samples were obtained for isotopic and BFAR analysis. These were washed over a 63 μ m sieve using water purified by reverse osmosis and dried in an oven at 50°C. Dry weights of the whole samples were recorded to calculate benthic accumulation rates. The coarse fraction >150 μ m was examined under a reflected light microscope and benthic foraminifera were isolated. The choice of species was restricted by assemblage composition. Species of Cibicidoides are widely used in paleoclimate studies because they are considered to include epifaunal species and appear to have consistent and predictable offsets from seawater "equilibrium values" [Shackleton and Hall, 1997; Katz et al., 2003a]. The abundance of individual species of Cibicidoides at Site 1218 varies considerably between samples and there were insufficient numbers of a single species in a consistent size range to permit a continuous monospecific record. To standardize analysis we generated records by picking two individuals each of C. grimsdalei, C. havanensis and a third common morphotype Cibicidoides praemundulus (termed C. subspiratus in the work by Coxall et al. [2005]; see Figure S1 in the auxiliary material) from the 150–300 μ m size fraction. In a few late Eocene samples, where calcitic microfossils are less abundant, this mix was modified to the specimens available (typically a greater weighting to C. grimsdalei).¹

[13] The *O. umbonatus* record was generated for two reasons: First, to evaluate the fidelity of the published *Cibicidoides* δ^{18} O and δ^{13} C records and second, as a means to

test for changes in δ^{13} C gradients across the sediment water interface that might shed light on deep sea organic carbon cycling across the EOT. *O. umbonatus* is more abundant than any individual species of *Cibicidoides* spp. and a monospecific record based on specimens from the 150–250 μ m size fraction was constructed. The δ^{18} O and δ^{13} C measurements were made on 304 samples of *Cibicidoides* spp. and 392 samples of the infaunal species *O. umbonatus*. *O. umbonatus* were taken from the same samples used in the study of *Coxall et al.* [2005] while the *Cibicidoides* spp. represent additional samples that double the resolution of the *Coxall et al.* [2005] record (now every 2.5 cm) through the earliest Oligocene.

[14] Isotopic analysis was performed at the National Oceanography Centre Southampton, University of Southampton, United Kingdom, using a Europa Geo 20-20 mass spectrometer equipped with a "CAPS" automatic carbonate preparation system. Results are reported relative to Vienna Peedee belemnite standard (VPDB). External analytical precision, based on replicated analyses of in-house standards calibrated to NBS-19 is $\pm 0.08\%$.

3.2. Site 1218 Eocene-Oligocene Chronology

[15] One of the primary benefits of the E-O section at 1218 E-O is the excellent age control afforded by the combination of well-defined magnetostratigraphic and biostratigraphic records and clear lithologic and geochemical cycles in cores, downhole logging data and geochemical proxies. Construction of a detailed stratigraphic correlation between Site 1218 and nearby Site 1219, provided opportunities to assess the "completeness" of the records, construct a quasi-continuous composite section and to constrain the position of magnetochron C13n within the XCB-cored interval of Site 1218. The latter was achieved by detailed stratigraphic mapping from Site 1219 (with an error of ± 10 cm), where the basal Oligocene magnetochron was recovered, to Site 1218 where it was not. The detailed chronology was generated by matching Site 1218 data with a back-projected astronomical model [Laskar et al., 2004; Pälike et al., 2005; Coxall et al., 2005]. The age model used here represents a refinement of the chronology

¹Auxiliary materials are available in the HTML. doi:10.1029/2010PA002021.



Figure 2. Revised ODP Site 1218 benthic foraminifera stable isotope records on the Site 1218 age scale; δ^{13} C with (a) n = 1089 and (b) n = 392 analyses. Solid lines are five-point running averages. Gray vertical arrow indicates double resolution in the *Cibicidoides* spp. record [after *Coxall et al.*, 2005]. "Zoom" arrow identifies the interval magnified in Figure 3. The two isotopic steps are designated EOT-1 and Oi-1 (see text).

used by *Coxall et al.* [2005] and involves an additional "automatic" tuning step [*Pälike et al.*, 2006]. On the Site 1218 time scale the estimated age of the E-O boundary is ~33.8 Ma (see Text S1 in the auxiliary material).

[16] On the Site 1218 age model, 5 cm sample spacing of *Coxall et al.* [2005], provides a maximum resolution of ~4 kyr. We have doubled the sampling density for the early Oligocene *Cibicidoides* spp. record, resulting in a theoretical maximum resolution of ~2 kyr in the early Oligocene where sedimentation rate is ~1–3 cm kyr⁻¹. While bioturbation acts to place an upper limit on the resolution achievable at sedimentation rates such as these, the higher fidelity of the new

epifaunal record compared to that of *Coxall et al.* [2005] bears testimony to the value of the higher-density sample set (section 4).

[17] So far there have been no attempts to produce a ³He normalized age model in the Paleogene EEP sediments. It is possible that this approach, which assumes a constant flux of extraterrestrial ³He (³He_{ET}) to the seafloor, might help further constrain fluxes [e.g., *Murphy et al.*, 2010], however, such methods come with their own uncertainties because of poor model calibrations that can overestimate sediment focusing [*Lyle et al.*, 2005]. Moreover, our cycle-based orbital age model yields a more finely resolved record than all previous

E-O accumulation rate-based Paleogene paleoproductivity reconstructions and thus goes a long way to minimizing chronological uncertainties at the 40 kyr cycle scale.

3.3. Benthic Foraminifer Accumulation Rate: A Paleoproductivity Proxy

[18] Changes in paleoproductivity are not straightforward to reconstruct from sediments because of variable preservation of biogenic components and chemical markers at the seafloor. Among the proxies available to Paleogene studies is the accumulation rate of deep-sea benthic foraminifer (BFAR). The BFAR proxy is based on the postulated relationship between surface primary production, organic carbon flux to the deep sea and the number of individual benthic foraminifers deposited [Herguera and Berger, 1991]. It must be applied carefully because taphonomic processes, especially calcite dissolution and bottom water oxygenation also influence assemblage composition [Herguera, 1992; Jorissen et al., 2007]. These considerations have hampered attempts to develop a quantitative proxy of organic matter flux but BFAR appears to correlate predictably with export production in a variety of modern locations [Herguera, 1992; Loubere, 1994; Jorissen et al., 2007] and the method has been used widely as a qualitative/semiquantitative proxy of paleoproductivity in the Pleistocene [e.g., Schmiedl and Mackensen, 1997; Gooday and Rathburn, 1999] and early Cenozoic [Smart, 2008; Alegret and Thomas, 2009]. The approach has been applied to E-O sequences from the Southern Ocean and the equatorial Atlantic [Diester-Haass and Zahn, 1996, 2001; Diester-Haass and Zachos, 2003; Schumacher and Lazarus, 2004] but only at low resolution in comparison to the isotope stratigraphies now available. Critical to accurate reconstructions of BFAR is a robust and detailed age model because changes in sedimentation rate, together with sediment density measurements, act as multipliers, and therefore have a large impact on calculated values of BFAR. The tight age control available at Site 1218 therefore, provides an unprecedented opportunity to apply the BFAR technique to the problem of EOT export production.

[19] Benthic foraminifera were counted independently in both the >250 μ m and the 150–250 μ m size fractions in 785 samples taken from the original *Coxall et al.* [2005] sample set. Two size fractions were counted to assess for preferential dissolution of smaller thinner walled specimens. The >250 μ m residues, which were typically <5 cm³, were counted in entirety while the >150 μ m samples were typically counted on a split (half) sample. Whole-sample dry weights were used to calculate the number of individuals per gram of dry sediment. BFAR was calculated following the methods of E. Thomas (personal communication, 2008) [after *Herguera and Berger*, 1991] by multiplying the number of benthic foraminifera per gram by the sample dry bulk density (DBD) and then by the linear sedimentation rate (LSR):

BFAR(number per cm² kyr⁻¹) = number of benthics per gram × DBD(g cm⁻³) × LSR(cm kyr⁻¹).

Constraints on DBD were derived by direct measurement on discrete samples and by regression from whole-core analyses of physical properties that are found to be highly correlated with measured values (Figure S2). Minor deviations in the physical properties data calibration may limit power to resolve the highest-frequency cyclic variations in sediment composition, however, the greatest source of uncertainty in BFARs propagate from the age model.

3.4. Cross-Correlation Analysis

[20] Cross-correlations of Site 1218 geochemical proxy records were carried out to explore lead-lag relationships between glaciation and carbon system changes. This method provides a measure of the similarity of waveforms as a function of a time lag applied to one of them and allows us to more confidently identify potential leads and lags during a fixed interval. The analysis was performed using Analys-Series 2.0.4.2 [*Paillard et al.*, 1996]. Data were evenly sampled before analysis. The time step used was 2.5 kyr for *Cibicidoides* spp. and bulk data pairs and 5.0 kyr for *O. umbonatus*, reflecting the maximum mean sampling resolution of data sets. Analysis was restricted to the second isotope shift of the transition (Figures 2 and 3), because it appears to be complete whereas the base of the first step is likely truncated by dissolution.

4. Results

4.1. Benthic Foraminifera Assemblages and Preservation

[21] The most common species in Site 1218 early Oligocene benthic foraminifer assemblages belong to the cosmopolitan, long-ranging genera Cibicidoides, Oridorsalis, Nuttallides, Gyroidinoides, Pullenia and Globocassidulina that are typical of lower bathyl to abyssal water depths above the CCD [Thomas, 1992; Katz et al., 2003b; Takata et al., 2007]. The scarcity of agglutinated species suggests welloxygenated conditions at the seafloor [Shipboard Scientific Party, 2002]. Detailed assemblage analysis was not part of this study. Takata et al. [2007] observed only minor changes in benthic assemblage composition in their study of early Oligocene sediments from Site 1218, however, their records are of low resolution and do not cover the critical E-O interval. Oligocene benthic foraminifera show signs of mild etching (Figure S1) on their shell surfaces but no evidence of severe corrosion, or shell breakage expected in response to extensive dissolution [Corliss and Honjo, 1981; Nguyen et al., 2009].

4.2. *Cibicidoides* spp. and *Oridorsalis umbonatus* Stable Isotope Records

[22] The most prominent features in the isotope records of *Coxall et al.* [2005], are a precursor hypothesized dissolution horizon and then two-step increase in δ^{18} O and δ^{13} C close to the E-O boundary with *Cibicidoides* spp. attaining maximum δ^{18} O and δ^{13} C in the basal Oligocene. The isotope series that we now present (Figure 2) substantiate this basic pattern, but their higher fidelity also brings to light important new structure.

[23] The two step pattern of benthic δ^{18} O step increase seen in our new infaunal record is very similar to that recorded in our improved epifaunal record with both series describing an initial increase of ~0.6‰, an intermediate plateau lasting ~300 kyr, followed by a further increase of ~0.9‰ in <40 kyr. The base of the first increase in δ^{18} O, here referred to as the EOT-1 for consistency with *Katz et al.* [2008], is not tightly constrained in our records because of the dissolution event



Figure 3. Analysis of climate proxy leads and lags across the Oi-1 isotopic step. (a) Close-up of Site 1218 benthic δ^{18} O, δ^{13} C, and bulk sediment weight %CaCO₃ between 33.59 and 33.75 Ma showing the timing of geochemical changes across Oi-1. (b) Cross-correlations of benthic δ^{18} O and δ^{13} C, bulk sediment δ^{18} O, and bulk %CaCO₃ as individual pairs for the Oi-1 interval. The *Cibicidoides* spp. and *O. umbonatus* records suggest that benthic δ^{13} C lags δ^{18} O by ~10 kyr across Step 2 whereas %CaCO₃ and δ^{18} O (all species) are in phase (see Table 2 for correlation statistics). Errors are equal to plus or minus half time step. Site 1218 bulk δ^{18} O is from *Pälike et al.* [2006].

Site	Location	Author	$\begin{array}{l} \text{Maximum} \\ \delta^{18} \text{O Shift} \\ (\text{\% VPDB}) \end{array}$	$\begin{array}{l} \text{Maximum} \\ \delta^{13}\text{C Shift} \\ (\text{\% VPDB}) \end{array}$	Species
522	South Atlantic	Zachos et al. [1996]	1.5	1.25	Cibicidoides spp. and Gyroidinoides sp.
689	subantarctic Atlantic	Diester-Haass and Zahn [1996]	1.5 ± 0.1	1	Cibicidoides spp.
744	Southern Indian Ocean	Zachos et al. [1996]	1.5 ± 0.1	1.1	Cibicidoides spp.
1218	Eastern equatorial Pacific	Coxall et al. [2005], this study	1.5 ± 0.1	1.2 ± 0.1	Cibicidoides spp. and O. umbonatus
1262	South Atlantic	Liu et al. [2004]	1.5	1.2 ± 0.1	Cibicidoides mundulus
1263	South Atlantic	Riesselman et al. [2007]	1.5 ± 0.1	1.2 ± 0.1	O. umbonatus
TDP12/17	Coastal Tanzania, W. Indian Ocean	Pearson et al. [2008]	1.5	0.8	T. ampliapertura (planktic)
	St Stephens Quarry, U.S. Gulf Coast	Katz et al. [2008]	1.2 ± 0.2		Cibicidoides pippeni

Table 1. The Magnitude of Deep Sea δ^{18} O and δ^{13} C Increases During the Eocene-Oligocene Transition Measured in Foraminifera From Locations Across the Globe^a

^aAll measurements are for benthics except *T. ampliapertura*, which is a mixed-layer planktonic species.

between 34 and 34.1 Ma. The second δ^{18} O step, referred to as Oi-1 and also after *Katz et al.* [2008], appears to be complete. Using the youngest Eocene samples before the dissolution horizon as a baseline, the overall increase to peak δ^{18} O in the earliest Oligocene has an amplitude of about 1.5‰ for *Cibicidoides* spp. and 1.6‰ for *O. umbonatus* equivalent to the global average as measured in different ocean basins (Table 1). The five-point moving averages plotted in Figure 2 show that superimposed on the intermediate plateau are two well-defined cycles of ~100 kyr duration each in both δ^{18} O series. Sedimentation rates in Site 1218 are low in the late Eocene (~0.5 cm kyr⁻¹), but our improved epifaunal record captures basic structure on the runup to the EOT with a particularly marked ~0.4‰ amplitude peak in δ^{18} O centered on ~34.37 Ma.

[24] Oligocene δ^{18} O reaches maximum values (~2.2‰ for *Cibicidoides* spp., 2.5‰ for *O. umbonatus*) around 33.65 Ma early in magnetochron C13n. The Oligocene record shows elevated δ^{18} O, on average 1.0‰ above the Eocene baseline, with marked, up to 0.5‰, δ^{18} O fluctuations. Phases of elevated δ^{18} O, comprising clusters of cycles are focused early in magnetochron C13n, upper C13n and around 32.68 Ma and this structure is more clearly defined in the higher-resolution epifaunal than in the infaunal record.

[25] Our new δ^{13} C records also substantiate the two step structure and amplitude of δ^{13} C increase described by the published records of *Coxall et al.* [2005]. The *O. umbonatus* record is more severely truncated by the precursor dissolution horizon than the epifaunal record making comparison for EOT-1 difficult to assess but the overall amplitude of increase from the late Eocene baseline to peak δ^{13} C in the earliest Oligocene is ~1.2‰ for *Cibicidoides* spp. and 1.0‰ for *O. umbonatus*. Following the abrupt increase in δ^{13} C there is a well-defined relaxation in both δ^{13} C series such that the overall pattern described is a characteristic "overshoot" [cf. *Zachos and Kump*, 2005]. In the δ^{18} O series the Oligocene baseline is distinctly offset to a new (higher) state than the Eocene baseline but for δ^{13} C, both data series return to nearlate Eocene values in the early Oligocene.

[26] Visual inspection of the isotope records reveals similar overall structure for δ^{18} O and δ^{13} C but there are some differences in phasing of these signals across the transition. Cross-correlation of carbon system proxies with δ^{18} O indicate that both the epifaunal and infaunal δ^{13} C records lag benthic and bulk carbonate δ^{18} O by 10 kyr (±1.25 kyr) on the build up to Oi-1 (Figure 3 and Table 2). In contrast %CaCO₃, *Cibicidoides* spp. δ^{18} O, and bulk sediment δ^{18} O are in phase.

[27] An obvious difference between our epifaunal and infaunal records is the large isotopic offsets with O. umbonatus registering higher δ^{18} O (by ~0.5‰) and lower δ^{13} C (by ~1.0‰). Calculation of epifuanal-infaunal $\Delta \delta^{13}$ C values during four phases of our Site 1218 Eocene-Oligocene interval reveals a small increase (by $\sim 0.4\%$) in the mean $\Delta \delta^{13}$ C value in samples after the Eocene dissolution event, and virtually no change in infaunal-epifaunal $\Delta \delta^{18}$ O (Figure 4), but there is no obvious correlation between changes in BFAR and $\Delta \delta^{13}$ C or $\Delta \delta^{18}$ O. The calculated offsets between species during the Oligocene, where the sample set is largest (n = 285), based on ordinary least squares linear regression are δ^{18} O Cibicidoides spp. = O. umbonatus -0.34 and δ^{13} C Cibicidoides. spp. = O. umbonatus +1.3 (Figure S3). These values are similar to those reported previously [Shackleton et al., 1984; Katz et al., 2003a; Billups et al., 2004], but they will likely yield over/under "corrections" because of the assumption that the slope equals 1 [Katz et al., 2003a].

4.3. Benthic Foraminifera and Carbonate Accumulation Rates

[28] Our BFAR records show a similar pattern in the >250 μ m and 150–250 μ m size fractions (Figure 5). For both, BFAR was low during the late Eocene but increased dramatically with CCD deepening. Early Oligocene BFAR shows a fluctuating pattern with high-frequency variation

Table 2. Cross-Correlation Statistics for Proxies During the Oi-1 Step of the E-O Isotopic Shift, 33.63–33.7 Ma

Parameter 1	Parameter 2	Sample Spacing (kyr)	Maximum Correlation Coefficient	Time Lag (kyr)
Cibicidoides δ^{13} C	Cibicidoides δ^{18} O	2.5	0.80	-10
Revised Cibicidoides $\delta^{18}O$	%CaCO ₃	2.5	0.89	0
<i>O. umbonatus</i> δ^{13} C	O. umbonatus δ^{18} O	5	0.85	-10
Bulk δ^{18} O	%CaCO ₃	2.5	0.91	0



Figure 4. Histograms showing the distribution of interspecies isotopic offsets during four phases of the EOT at Site 1218. (a and b) Early Oligocene postglacial maximum (33.05–32.52 Ma). (c and d) First 600 kyr of the Oligocene including Oi-1 (33.67–33.05 Ma). (e and f) EOT pre-Oi-1 (33.92–33.67 Ma). (g and h) Eocene predissolution event (35.21–34.18 Ma). $\Delta \delta^{13}$ C is *Cibicidoides* spp. minus *O. umbonatus* δ^{13} C; $\Delta \delta^{18}$ O is *O. umbonatus* minus *Cibicidoides* spp. δ^{18} O.



Figure 5. Site 1218 benthic foraminifera and bulk CaCO₃ flux constraints. (a and b) Abundance of benthic foraminifera in the >250 μ m and 150–250 μ m size fractions. (c) Bulk sediment weight %CaCO₃ measured (black) and estimated (gray) (based on multisensor-track (MST) color reflectance L* calibration) [*Coxall et al.*, 2005]. (d) Linear sedimentation rate (LSR) based on the Site 1218 age model. (e) Sediment dry bulk density estimated from multisensor-track gamma ray attenuation porosity determination bulk density (DBD) measurements [*Pälike et al.*, 2006] (see also Figure S2). (f) CaCO₃ accumulation rate (AR) [after *Coxall et al.*, 2005]; (g and h) Benthic foraminifer accumulation rate (BFAR) calculated for the two size fractions. Thick lines represent five-point running averages.

and an interval of elevated values between 33.4 and 33.8 Ma. The data describe two separate peaks of >3000 BF cm² kyr⁻¹ in the 150–250 μ m fraction and ~2000 BF cm² kyr⁻¹ in the >250 μ m fraction, falling to <200 BF cm² kyr⁻¹ in between. These maxima are around 2 to 3 times greater than modal Oligocene values (500–600 BF cm² kyr⁻¹ for both fractions), with the extremes reaching 4 to 5 times the mode. From ~33.35 Ma, the BFAR records decrease to a low of ~100 BF cm² kyr⁻¹, fluctuating between 100 and 600 BF cm² kyr⁻¹. This pattern is interrupted at 32.9–33.1 Ma and 31.4–31.6 Ma, when mean BFAR values increase to modes of about 1200 and 900 BF cm² kyr⁻¹ in the 150–250 μ m and >250 μ m fractions, respectively.

[29] Early Oligocene bulk sediment CaCO₃ accumulation rate (CaCO₃ AR) increased abruptly at \sim 33.70 Ma to a maximum (overshoot) of >1.0 g cm⁻³ kyr⁻¹ in the basal Oligocene. Elevated values 2 to 3 times the Oligocene base line of \sim 0.4 g cm⁻³ kyr⁻¹ were maintained over two pulses before decreasing at ~33.36 Ma to a ~40 kyr long minimum (Figure 5f). The BFAR increase that is coincident with the CCD shift is driven by high accumulation rate estimates. Subsequently, $CaCO_3$ AR increased again, settling into a pattern of strong, high-frequency fluctuations with similar amplitude variation to the earliest Oligocene maximum.

5. Discussion

5.1. Structure of the Eocene-Oligocene Glaciation

[30] Our new Site 1218 climate proxy records provide the most detailed view to date of the structure of the EOT from the World's largest ocean basin, confirming its two-step structure at Site 1218. In Figure 6 we compare our new isotope records with published data sets from other drill sites. This exercise serves to strengthen the case for an EOT precursor late Eocene ice growth event [cf. *Katz et al.*, 2008; *Peters et al.*, 2010; *Pusz et al.*, 2011] but better resolved records are needed from the Late Eocene to determine whether or not this



Figure 6. Global deep sea δ^{18} O and δ^{13} C comparisons for Site 1218 (this study), Site 689 (sub-Antarctic Atlantic) [*Diester-Haass and Zahn*, 1996], Site 522 (South Atlantic) [*Zachos et al.*, 1996], and Site 744 (southern Indian Ocean) [*Zachos et al.*, 1996] (see Figure 1). Approximate conversions to the Site 1218 timescale were made for each site by aligning common magneto-stratigraphic tie points. Benthic data are all *Cibicidoides* spp. and have been adjusted to equilibrium values (+0.64 added to δ^{18} O) [*Shackleton and Hall*, 1997]. Oi-1a and Oi-1b δ^{18} O events follow the nomenclature and concept of *Zachos et al.* [1996]. Gray shading represents possible correlation of these events between sites. In reality, they should be synchronous, and the obvious differences in duration indicated here likely result from age model inconsistencies and the crude attempt at conversion made here.

transient represents one in a series of comparatively lowamplitude cycles in δ^{18} O. Regardless, our records show that any glacial advances that might have occurred on Antarctica on the immediate runup to the EOT must have been comparatively small and transient in nature and that large ice sheets were not sustained until the EOT itself. Our records also emphasize that these large ice sheets must have grown extremely rapidly in keeping with the hypothesized role of powerful nonlinear feedbacks in ice sheet mass balance [Weertman, 1976; Oerlemans, 2002; DeConto and Pollard, 2003; Pollard and DeConto, 2005; Coxall et al., 2005]. Our new records reveal more variance during the interval between the two pronounced δ^{18} O steps than seen in the records of Coxall et al. [2005] thereby serving to strengthen correspondence between Site 1218 data and the results of ice sheet coupled climate model experiments [DeConto and Pollard, 2003]. Our results do not, however, strengthen the case for a third δ^{18} O step, indicating that "EOT-2" interpreted from the U.S. Gulf Coast shelf record [Katz et al., 2008] corresponds to an interval of amplitude variation change not seen in the deep sea records.

[31] Based on δ^{18} O records from the South Atlantic and Southern Ocean [Miller et al., 1991; Zachos et al., 1996] a case has been made for a "zone" of maximum δ^{18} O that represents an early Oligocene glacial maximum. This has been referred to as Oi-1 (also the Early Oligocene Glacial Maximum) [Liu et al., 2004; Coxall and Pearson, 2007], modifying the original concept of Oi-1 as an interval of abrupt δ^{18} O increase (and applied here) as defined by *Miller* et al. [1991] for lower-resolution records. In fact, the four records presented in Figure 6 lend strong support to the concept of two distinct glacial maxima within magnetochron C13n Oi-1a and Oi-1b [Zachos et al., 1996] and both of these maxima are associated with peaks in δ^{13} C that lag δ^{18} O by about 10 kyr (Figure 7b). Nevertheless, the main conclusion to be drawn from the new Site 1218 record is that the EOT signal cannot really be adequately understood as a sequence of discrete "events," because it is clearly bears the imprint throughout of strong variability at the 10–100 kyr (orbital) scale (Figure 6).

[32] Oi-1a and Oi-1b glacial maxima appear to be less well defined in our infaunal than our epifaunal record (Figure 3), a pattern also seen in records from Walvis Ridge, South Atlantic (*Zachos et al.* [1996], Site 522; *Riesselman et al.* [2007], Site 1262), although δ^{13} C shows two well-defined maxima. The cause of this is unknown but may reflect a habitat difference.

5.2. CCD Deepening

[33] Cross correlation analysis confirms that the deep sea and near-surface (bulk) δ^{18} O increase are in lock step with increasing carbonate preservation over Oi-1 (Figure 3). This finding is consistent with the suggestion [*Coxall et al.*, 2005] that CCD deepening was closely associated with Antarctic glaciation. One potential mechanism for explaining the link between the onset of sustained Antarctic glaciation and the basification of the deep Pacific Ocean across the EOT is a shelf-to-basin shift in marine carbonate sedimentation in response to glacioeustatic sea level fall [*Coxall et al.*, 2005]. The results of nutrient capable geochemical box model experiments have lent support to this suggestion [*Merico* *et al.*, 2008]. In these experiments shelf basin fractionation achieves the observed temporary increase in seawater δ^{13} C and "overshoot" in CCD deepening by a one off "dump" into the global ocean of isotopically "heavy" weathered carbon from newly exposed shelf carbonates while the permanent shift to a deeper CCD state is induced by the requirement of the ocean to maintain the input/output balance of carbonate in response to the reduction in shelf habitat for neritic calcifiers [cf. *Zeebe and Westbroek*, 2003; *Ridgwell and Zeebe*, 2005]. This explanation, however, has been called into question on a number of grounds.

[34] One question concerns the glacioeustatic history of the EOT. The shelf basin fractionation model for explaining EOT CCD deepening has been criticized as being incompatible with the best available sequence stratigraphic reconstructions of sea level change across the EOT [Rea and Lyle, 2005; Miller et al., 2009]. Yet the considerable geophysical uncertainties associated with the sequence stratigraphic method on even "passive" continental margins means that more globally distributed records of this type are needed before we can confidently extract the eustatic signal [Moucha et al., 2008]. Similarly, attempts to separate the ice volume (eustacy) and temperature change components from the benthic δ^{18} O signal using Mg/Ca paleothermometry have met with difficulties because of the competing effects of changes in deep sea $[CO_3]^{2-}$ chemistry that appears to mask the temperature effect in Mg/Ca [Lear et al., 2004, 2008, 2010; Lear and Rosenthal, 2006; Coxall et al., 2005; Yu et al., 2007; Pusz et al., 2011]. Where this problem has been circumvented in shallow shelf sequences, results indicate that the overall δ^{18} O signal is attributable to both cooling and ice growth with increasing contribution of ice volume as the EOT progressed [Lear et al., 2008, 2010; Katz et al., 2008; Pearson et al., 2008]. At Site 1218, however, the records suggest tight coupling of δ^{18} O, CaCO₃ and lagged δ^{13} C for both EOT-1 and Oi-1, implying that the two δ^{18} O steps share a common origin. Increases in benthic Li/Ca at Site 1218 [Lear and Rosenthal, 2006; Lear et al., 2010] and in the south Atlantic (ODP Site 1263) [Peck et al., 2010], which are interpreted to indicate significant increase in $[CO_3]^{2-}$ associated with both steps, is consistent with this hypothesis.

[35] A second question that has been raised over the ability of shelf basin fractionation to account for the EOT stems from a number of studies [Salamy and Zachos, 1999; Latimer and Filippelli, 2002; Anderson and Delaney, 2005; Ravizza and Paquay, 2008; Dunkley Jones et al., 2008] that suggest that the EOT increase in δ^{13} C was caused by an increase in global organic carbon burial (output of ¹²C-rich carbon, not input of ¹³C-rich carbon) perhaps coupled with onset of glacial weathering/weatherability [Zachos and Kump, 2005]. Merico et al. [2008] examined mechanisms by which increased C_{org} burial might be achieved (more vigorous ocean circulation in a cooler climate, increased efficiency of organic carbon burial at the seafloor, and increased nutrient delivery to the oceans) but none faired well (see also section 5.4).

5.3. Eastern Equatorial Pacific Productivity Change

[36] Our BFAR record provides evidence for increased numbers of benthic foraminifera in the EEP associated with the EOT. Depth-dependent preservation of calcite has undoubtedly influenced the sedimentary record of calcite



Figure 7. Carbon system changes. (a) Site 1218 BFAR >150 μ m (150–250 μ m and >250 μ m data combined). (b) Site 744 opal accumulation [*Salamy and Zachos*, 1999]. Site 1218 *Cibicidoides* spp. (c) δ^{13} C and (d) δ^{18} O. (e) Tropical Indian Ocean atmospheric pCO₂ proxy record derived from boron isotopes plantkonic foraminifera from Tanzania [*Pearson et al.*, 2009]. Red squares are reconstructed pCO₂ for variable carbonate ion concentration ([CO₃]²⁻), and green diamonds are constant [CO₃]²⁻. (f) Site 1218 bulk %CaCO₃. (g) CaCO₃ accumulation rate (AR) [after *Coxall et al.*, 2005]. Figures 7a–7d are five-point running averages. All data are plotted on the Site 1218 age scale. Conversion of the Site 744 data set to the Site 1218 time scale is approximate (based on existing stratigraphic tie points) (S. Bohaty, personal communication, 2010).

deposition at Site 1218 and so before we attempt to interpret the BFAR record we must first separate the CCD signal.

[37] For the most soluble sedimentary carbonate fraction (planktonic foraminifera and calcareous nannofossils), the relationship between the CaCO₃ sediment dissolved and that remaining is highly nonlinear such that, when the bulk CaCO₃ fraction is large, small percentage loss in bulk CaCO₃ indicates substantial CaCO₃ dissolution [*Berger*, 1971; *Thunell et al.*, 1981; *Broecker*, 1995; *Stap et al.*, 2009]. Oligocene variations in bulk carbonate accumulation at Site 1218, therefore, are likely responses to fluctuations in lysocline depth even once the CCD had deepened [*Pälike et al.*, 2006]. In contrast, the tests of hyaline deep sea benthic foraminifera, which live and calcify at depths as great as 4000 m post-Eocene and are much more refractory than bulk carbonates, typically showing little evidence of dissolution in carbonate-rich sediments [*Berger*, 1971; *Corliss and Honjo*, 1981; *Thunell et al.*, 1981; *Nguyen et al.*, 2009]. Thus, while preservation is the dominant control on BFAR in the latest Eocene, once the CCD had deepened, such that bulk sediment CaCO₃ at Site 1218 increased to ~75% and benthic foraminifera show minimal signs of dissolution, we interpret our BFAR record to represent an abundance signal. This interpretation, and >75% criterion, is supported by nannoflora preservation studies [*Blaj et al.*, 2009] that record full Oligocene diversity in samples with >75% CaCO₃, and by the similarity in form of the BFAR records generated on both the smaller (150–250 μ m) and larger (>250 μ m) size fractions.

[38] Restricting interpretation to the interval of >75% CaCO₃, which occurs from ~ 33.7 Ma (equivalent to the Oi-1 step) onward, our record suggests a transient (~ 500 kyr) twofold to threefold increase in BFAR relative to baseline Oligocene values during magnetochron C13n, in two principal pulses (Figures 5 and 7). We interpret these peaks to represent increased surface biological production and/or increased efficiency of organic carbon export in the EEP associated with early Oligocene Antarctic glaciation. A third more modest increase in BFAR at around 33 Ma suggests a subsequent pulse after peak glacial conditions.

[39] The δ^{13} C differences between epifaunal and infaunal benthic foraminiferal species have been used to evaluate organic matter remineralization at the seafloor, with high $\Delta \delta^{13}$ C corresponding to high organic matter remineralization rates at times of maximum organic fluxes [e.g., Schmiedl and Mackensen, 2006; Jorissen et al., 2007]. The small (0.4‰) offset to higher $\Delta \delta^{13}$ C in the early Oligocene relative to the late Eocene (Figure 4), therefore, may be a consequence of the early Oligocene EEP productivity changes. In the Eocene, Corg may have been more efficiently recycled above the sediment water interface whereas in the Oligocene, overall more C_{org} made it into pore waters to be remineralized there. Decreasing ocean temperature associated with the EOT could have had a role in this process [Olivarez Lyle and Lyle, 2006]. The lack of correspondence between $\Delta \delta^{13}$ C and maximum BFAR may be related to the shallow infaunal habitat of O. umbonatus that may not capture the full magnitude of pore water DIC signal. Additional isotope data from a deep infaunal species, and expansion of the assemblage-based data set after *Takata et al.* [2007] might provide additional insights into this question.

[40] Records of opal accumulation rate in the Southern Indian Ocean (ODP Site 744) [Salamy and Zachos, 1999] demonstrate a marked increase in the early Oligocene (Figure 7). Controlled partially by an increase in diatom accumulation [Salamy and Zachos, 1999], the increase in opal suggests the initiation of significant seasonal production in the Southern Ocean facilitated by deep mixing and upwelling supplying silica to the photic zone [Salamy and Zachos, 1999; Berger, 2007]. In this context the modest pulse in diatom abundance that occurs in the early Oligocene at Site 1218 and several other equatorial Pacific sites may take on added significance [Fenner, 1984; Baldauf, 1992; Barron et al., 2004; J. Bauldauf, unpublished data, 2010].

[41] On the other hand, low-resolution barite and CaCO₃ accumulation records from several equatorial Pacific ODP Sites, including Site 1218, have been used to argue against increased equatorial productivity during the E-O, instead suggesting a decrease in the rain ratio of organic to inorganic carbon from the late Eocene to early Oligocene [Schumacher and Lazarus, 2004, Griffith et al., 2010]. Interpreting organic export accumulation at E-O transition time, however, is complicated by major changes in the processes of ocean biomass production, transport, preservation and storage, which, like calcite preservation, appear to have undergone fundamental state changes across the EOT. Such changes include the possible influence of decreased ocean temperature on the production and preservation of organic carbon [Olivarez Lyle and Lyle, 2006] and a hypothesized increase in the dependence of deep sea benthos on surface supplied food (development of bentho-pelagic coupling), compared to in situ and laterally transported food, which are considered to have been important for Mesozoic and early Cenozoic benthos [*Thomas and Gooday*, 1996]. Increases in diversity and mean size of siliceous phytoplankton, especially diatoms [*Falkowski et al.*, 2004; *Finkel et al.*, 2005] at this time, may have also played a role in global patterns of export production and preservation of biogenic remains, although the effects of reduced ballasting associated with a shift toward siliceous primary production at this time is unknown [e.g., *Francois et al.*, 2002; *Iversen and Ploug*, 2010]. The potential impact of these factors on C_{org} distribution and preservation in the deep sea equatorial Pacific across the EOT, are still largely unconstrained [*Archer et al.*, 1993; *Olivarez Lyle and Lyle*, 2006].

[42] On the balance of evidence it seems that E-O climate change did not trigger a major state shift toward increased production in the EEP, but there is evidence for a short-lived increase in productivity and export production associated with the onset of sustained Antarctic glaciation. Furthermore, the first-order similarities between our BFAR record from the Equatorial Pacific Ocean and the Southern Ocean record of opal accumulation suggests that the productivity increases at the two sites are in some way linked. One potential mechanism to explain this inferred high-to-low-latitude link is through the development of deep mixing in the Southern Ocean, strengthening of ocean ventilation and the development of a common (nutrient/silica-rich) chemistry in waters outcropping at the surface at both Site 744 and 1218, as recognized over more recent times scales [Toggweiler and Carson, 1995; Dugdale et al., 2002; Loubere et al., 2003; Sarmiento et al., 2004]. This is supported by sedimentary and geochemical data that indicates an intensification of thermohaline circulation in the early Oligocene including increased influence of Southern Ocean deepwater in the tropical Pacific [Miller and Tucholke, 1983; Wright and Miller, 1993; Davies et al., 2001; Scher and Martin, 2008; van de Flierdt et al., 2004; Via and Thomas, 2006; Thomas et al., 2008; Cramer et al., 2009; Pusz et al., 2011].

[43] A second possibility is that the inferred productivity teleconnection between these two sites is driven by increased atmospheric circulation, that is, increased strength of trade winds [*Moore et al.*, 2002; *Huber*, 2002; *Lyle et al.*, 2008], and a focusing of equatorial sedimentation attributable to changes in the pattern of equatorial divergence [*Lyle*, 2003; *Moore et al.*, 2004, 2008; *Vanden Berg and Jarrard*, 2004; *Rea and Lyle*, 2005]. In reality, as in the modern, productivity variation in the equatorial Pacific is probably controlled by multiple factors [*Loubere et al.*, 2003; *Anderson et al.*, 2009]. Our ability to test these hypotheses is currently limited by the lack of detailed spatial reconstructions of E-O sediment flux patterns and the complications of changes in carbonate preservation associated with CCD deepening.

5.4. The Cause of the δ^{13} C Excursion

[44] The prominent excursion in δ^{13} C, extending 400– 500 kyr into the early Oligocene, indicates a major perturbation of the global carbon system associated with the onset of Antarctic glaciation but the cause of this perturbation, as well as the link with CCD deepening, is debated. Arguably, the most obvious explanation for the >1.0‰ δ^{13} C increase is a global increase in export and burial of organic carbon in the deep sea. Our BFAR records, together with records of opal burial in the Southern Ocean provide some support for a productivity increase (section 5.2). However, our deep-sea records appear to be symptomatic of C_{org} export to the seafloor but not carbon sequestration (burial) because Paleogene EEP sediments exhibit C_{org} contents of <0.1 wt% (an order of magnitude lower than Holocene EEP values) [*Lyle et al.*, 2002; *Olivarez Lyle and Lyle*, 2006; *Pälike et al.*, 2009]. Elsewhere, for example, the Paratethys, southern Indian and west African margins, C_{org} exported was not remineralized but buried [*Schulz et al.*, 2002; *Sarkar et al.*, 2003; *Ravizza and Paquay*, 2008]. The "missing carbon," therefore, may be accounted for in regions such as these that are close to continents, and where 90% of C_{org} burial takes place today [*Berner and Canfield*, 1989; *Hedges and Keil*, 1995]. Regardless, the processes that result in such low C_{org} in the Equatorial Pacific require further study.

[45] A second problem associated with the C_{org} burial hypothesis is that, although geochemical modeling studies can simulate the $\delta^{13}C$ excursion by increased organic carbon burial, thus far they have been unable to simulate the observed CCD response by the same mechanism [*Zachos and Kump*, 2005; *Merico et al.*, 2008]. In contrast, simulations forced by sea level–led increases in weathering inputs of isotopically heavy neritic limestone reproduced both a permanent CCD deepening and a transient $\delta^{13}C$ perturbation over comparable time scales to that observed [*Merico et al.*, 2008]. That said, the way in which the shelf basin fractionation hypothesis is instantiated into the modeling of *Merico et al.* [2008] needs to be improved to consider C_{org} (not just CaCO₃) fractionation.

[46] One critical question to establish is the extent to which δ^{13} C increase and CCD deepening are forced by the same mechanism. Our detailed records reveal cycle-level similarities in the form of δ^{13} C, δ^{18} O and %CaCO₃ but there they are not synchronous. The ~10 kyr lag of δ^{13} C with respect to δ^{18} O and %CaCO₃ over Oi-1 (Figure 5), which has been identified at several sites [Zachos et al., 1996; Riesselman et al., 2007], implies an indirect or delayed response of the driver of δ^{13} C change. One explanation is that δ^{13} C and δ^{18} O were forced by the same climatic mechanism, for example, ice sheet growth (shelf to basin transfer), but the long residence time of carbon in the ocean sediment reservoir (~100 kyr) [Broecker and Peng, 1982] created a delay in benthic δ^{13} C change [*Pälike et al.*, 2006]. Alternatively, if the δ^{13} C increase was associated with changes in biomass production and burial the delay might signal a "nutrient lag" reflecting shelf to basin delivery of biologically important nutrients such as phosphorous [Broecker, 1982; Filippelli et al., 2007]. Moreover, the carbon system likely responded to different periodic forcing mechanisms at different times during the EOT, because δ^{13} C and δ^{18} O appear to be synchronous over the highest-frequency cycles during the E-O intermediate plateau (Figure 2), yet the spectral analyses of Coxall et al. [2005] show that δ^{13} C leads δ^{18} O (by ~8 kyr) in the 40 kyr band from ~33.5 Ma onward.

5.5. Eocene-Oligocene Glaciation and Atmospheric Carbon Dioxide

[47] The role of atmospheric CO_2 in forcing and/or responding to large-scale glaciation at the EOT is at the heart of trying to understand the onset of Antarctic glaciation. Changes in tectonic CO_2 sources and global weathering patterns are thought to be responsible for the long-term pattern of decline in Paleogene CO₂ proxies records [*Pearson and Palmer*, 2000; *Pagani et al.*, 2005]. It seems unlikely that increases in global export production and burial was the cause of Antarctic glaciation [*Zachos and Kump*, 2005; *Merico et al.*, 2008], however enhanced productivity in the early Oligocene may have provided a positive feedback role by triggering further more rapid pCO₂ decline [*Zachos et al.*, 1996; *Diester-Haass*, 1995, *Dalai et al.*, 2006; *Ravizza and Paquay*, 2008; *DeConto et al.*, 2008] that promoted glacial expansion.

[48] The most recently published pCO₂ proxy record for the EOT, which is based on boron isotopes in well-preserved planktonic foraminifera from Tanzania [*Pearson et al.*, 2009], is consistent with the basic idea that initiation of Antarctic glaciation was associated with pCO₂ decline during the late Eocene (Figure 7). This decline (~300 ppm from 34.5 Ma to 33.7 Ma), which is essentially constrained by a single data point, is followed by a temporary reversal to late Eocene values roughly in step with rising δ^{18} O. pCO₂ subsequently decreased to a new lower Oligocene baseline after peak glacial conditions were initiated, a pattern implied by longer-term records [*Pearson and Palmer*, 2000; *Pagani et al.*, 2005].

[49] The data suggest that Antarctic glaciation was closely associated with a temporary increase in pCO₂. Age control in the Tanzanian records is weak in comparison to our records but if the structure and timing of this pCO₂ increase is substantiated and shown to be globally representative then it is fundamentally inconsistent with a Corg-burial-led explanation for δ^{13} C increase. There is some suggestion based on nannofossil that the lower Oligocene Tanzanian sequence records evidence of eutrophication [Dunkley Jones et al., 2008] suggesting that the site is unlikely to have been at air-sea equilibrium with respect to CO₂. Yet, the magnitude of local ocean-atmosphere disequilibrium required to explain the observed signal in this way is large. One alternative to consider is that the CO₂ increase was driven by ocean "venting" if, as proxies suggest, overturning circulation increased during the early Oligocene, a mechanism that (1) has been proposed to help explain increases in pCO_2 at the end of the last glacial cycle [Anderson et al., 2009; Skinner et al., 2010] and (2) yields a short-lived ventilation spike in CO_2 in the model results of *Merico et al.* [2008]. The residence time of CO_2 in the subsurface ocean, however, is short (~1000 years) compared to geological time scales and is unlikely to have caused a 300-400 kyr long pCO₂ excursion suggested by the proxy record. Other possible explanations for a CO₂ increase at the EOT are a reduction in silicate weathering [Broecker and Peng, 1982; Zachos et al., 1999; Zachos and Kump, 2005; Lear et al., 2004; Vance et al., 2009], or a release of organic carbon from terrestrial or continental shelf reservoirs [Pearson et al., 2009].

6. Conclusions

[50] We present new deep sea paleoceanographic proxy records from eastern equatorial Pacific ODP Site 1218 that provide insight into the role of the ocean carbon system in Eocene-Oligocene climate change.

[51] 1. Our new isotope records support a rapid two-step structure of δ^{18} O increase across the EOT described in the records of *Coxall et al.* [2005] but reveal more variability in

the intervening "plateau" interval. This result serves to only increase the convergence between the Site 1218 data series and the model reconstructions of *DeConto and Pollard* [2003], consistent with threshold crossing behavior and the operation of powerful positive feedbacks associated with ice sheet height and coalescence.

[52] 2. Integration with global deep-sea records provides new support for the hypothesized existence of a precursor Late Eocene δ^{18} O increase event [*Katz et al.*, 2008] on the runup to the onset of full blown Antarctic glaciation and for two early Oligocene glacial maxima (Oi-1a and Oi-1b) [*Zachos et al.*, 1996]. Nevertheless, the improved fidelity afforded by our new records emphasizes that the EOT cannot be adequately understood as a series of discrete events because it is clearly imprinted by orbitally paced variability throughout.

[53] 3. Our record of BFAR record is the best resolved data set of its type available for this time interval and unique in being the first to include the effects of orbital-scale changes in sedimentation rate and sediment density. BFAR appears to reveal temporary (~500 kyr), pulsed, increase in export production in the EEP associated with maximum Oligocene glacial conditions. First-order similarities in structure to Southern Ocean opal records suggests strengthening of a connection between high and low latitudes through ocean circulation, effecting chemistry of water upwelling to the EEP thermocline. Intensification of equatorial winds associated with E-O climate change could also have played a role.

[54] 4. The offset in δ^{13} C between our epifaunal and shallow infaunal benthic foraminifera are higher in the early Oligocene than late Eocene (by ~0.4‰), consistent with an overall increase in sediment flux and seafloor remineralization but there is no correspondence to individual BFAR peaks, perhaps reflecting the shallow infaunal habitat of *O. umbonatus*.

[55] 5. The inferred increase in biological pumping and well-documented increase in benthic δ^{13} C are not readily reconciled with the timing of the transient pCO₂ rise observed in a recent proxy reconstruction of the E-O transition from Tanzania [*Pearson et al.*, 2009].

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