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Astronomic and Oceanographic Influences on Global Carbon Cycling across the Oligocene/Miocene Boundary

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Introduction

Oligocene/Miocene organic carbon burial

The $\delta^{13}C$ composition of marine inorganic carbon on geologic time scales has been used to assess the balance between burial and weathering of organic and inorganic carbon (e.g., Shackleton, 1984, 1985; Kump and Garrels, 1986). This is based on the observation that storage of light organic carbon in marine sedimentary reservoirs (~20 to ~23\%e) or continental reservoirs (~25\%e) leaves the residual marine dissolved inorganic carbon (DIC) pool enriched in $\delta^{13}C$ (e.g., Broecker and Peng, 1982). For example, a sustained increase in marine organic carbon burial of $2.1 \times 10^{12}$ tons C would enrich ocean $\delta^{13}C$ by ~1\%e (e.g., Vincent and Berger, 1985). Shackleton (1985) produced a Cenozoic $\delta^{13}C$ record based on bulk carbonate sediment that documented four major episodes of increased relative organic carbon burial, during the middle Eocene, early Oligocene, Oligocene/Miocene (O/M) boundary, and early to middle Miocene. These increases are also recorded in foraminiferal calcite, although other processes modify their $\delta^{13}C$ composition.

A distinct positive excursion in benthic and planktic foraminiferal $\delta^{13}C$ of ~1\%e was confirmed near the O/M boundary (Miller et al., 1991a; Wright et al., 1992; Hodell and Woodruff, 1994; Flower et al., 1997a, b) ca. 23 Ma on the timescale of Cande and Kent, 1992, 1995). The excursion began near the top of Subchron C6Cr and ended near the Chron C6AA/C6Bn boundary (Hodell and Woodruff, 1994). In fact, this $\delta^{13}C$ maximum has since been utilized as a distinct stratigraphic marker for the O/M boundary (Hodell and Woodruff, 1994). The $\delta^{13}C$ excursion is composed of several individual carbon isotope maxima that have a periodicity of approximately 400 k.y. (Zachos et al., 1997; 2001b, Paul et al., 2000). Carbonate preservation is good during the maxima and poor in between, suggesting control by ocean circulation and chemistry changes. A similar relationship occurred during the “Monterey Carbon Isotope Excursion” in the early to middle Miocene (Vincent and Killingley, 1985; Vincent and Berger, 1985; Miller et al., 1991a, b; Woodruff and Savin, 1991; Flower et al. 1993a, b).
Trade wind induced upwelling

One hypothesis for increased organic carbon burial near the O/M boundary is essentially an application of the “Monterey Hypothesis” originally developed for the early to middle Miocene δ¹³C excursion (Vincent and Berger, 1985). The “Monterey Hypothesis” holds that polar cooling increased the planetary temperature gradient, which increased trade wind intensity and in turn upwelling, particularly in coastal regions around the Pacific marked by marginal basins (Vincent and Berger, 1985). A similar mechanism may have operated near the O/M boundary. High-resolution stable isotope records demonstrate that δ¹³C maxima within the excursion are associated with δ¹⁸O maxima, confirming a relation to deep sea cooling and ice growth (Flower et al., 1997a; Zachos et al., 1997). A key prediction of this mechanism is increases in trade wind strength, at least during the individual δ¹³C maxima. Such increases may be necessary to enhance coastal upwelling and associated organic carbon deposition. However, there is as yet no evidence for trade wind strength increases during individual δ¹³C maxima (Hovan and Rea, 1992). Indeed, eolian dust flux near the O/M boundary is not particularly high relative to the rest of the Cenozoic (Ree, 1994).

More recently, proxy estimates of latest Oligocene to Miocene pCO₂ indicated surprisingly low levels throughout the Miocene (Pagani et al., 1999; Pearson and Palmer, 2000). Although these data hint at orbital-scale variations in Miocene pCO₂, mean levels (200-300 ppmv) appear similar to preindustrial levels. Thus it is possible that the O/M boundary and early to middle Miocene δ¹³C excursions were closely associated with atmospheric CO₂ drawdown (Pagani et al., 1999; Flower, 1999; Pearson and Palmer, 2000).

Nutrient delivery

Raymo (1994) advanced the hypothesis that Miocene organic carbon deposition increased because of greater nutrient delivery to the oceans related to Himalayan uplift. This hypothesis does not invoke trade wind variability. She used the strontium isotope composition of marine carbonate (Richter et al., 1992) as a proxy of continental weathering, and in combination with Shackleton’s (1985) estimate of organic carbon vs. carbonate burial, suggested a pronounced increase in organic carbon burial during the early to middle Miocene (Raymo, 1994). Increased continental nutrient supply could also have been derived from other regions (Ravizza, 1993; Raymo, 1994; Reusch et al., 1998). The nutrient delivery hypothesis is consistent with the observation that carbonate accumulation also increased during the mid-Miocene (Woodruff and Savin, 1991), although this trend could alternatively be due to regional shifts in carbonate preservation. This model can be applied to the O/M boundary. Increased nutrient supply should enhance the “biologic pump” removing nutrients from surface waters and returning them to sub-thermocline waters via remineralization (e.g., Broecker and Peng, 1982). Such remineralization should also enhance development of the oxygen minimum zone (OMZ) in certain areas.

Sea-level-controlled upwelling

One other hypothesis that deserves mention is sea-level-controlled upwelling on shallow carbonate banks (Compton et al., 1990). Widespread phosphorite deposits in the southeastern U.S. have been associated with δ¹³C maxima from the O/M boundary to the middle Miocene (Compton et al., 1990; 1993; Mallinson and Compton, 1997). Originally rich in organic carbon, these deposits are inferred to have contributed to a mean ocean δ¹³C increase, and subsequently diagenetically altered to phosphorite. Noting that these deposits roughly coincide with Miocene δ¹³C maxima during an interval of generally rising sea level on the >1 m.y. scale, these workers proposed that topographically induced upwelling in this region may have resulted in organic carbon and phosphate accumulation. However, the association of δ¹³C maxima with δ¹⁸O maxima (sea-level lowering) on the <1 m.y. scale indicates
that different processes control organic carbon burial on orbital time scales (Mallinson and Compton, 1997). Nevertheless, it is possible that sea-level-controlled upwelling played a significant role in organic carbon burial on the super-orbital scale during the O/M $\delta^{13}$C maximum.

**Approach**

**Western Equatorial Atlantic hydrography as a monitor of global change**

A central tenet of the Monterey Hypothesis is that increased planetary thermal gradients enhanced trade wind strength, and hence coastal upwelling (Vincent and Berger, 1985). Increased trade winds would also have two important effects in equatorial settings: (1) Open ocean upwelling would increase in central and eastern regions, and (2) warm surface waters would "pile up" in western equatorial areas, leading to a thicker mixed layer (and a deeper thermocline; Merle, 1980). These effects should have been partially reduced in the Miocene ocean because the Isthmus of Panama was open to at least intermediate depths (Duque-Caro *et al.*, 1990) and therefore the bathymetric barrier that allowed piling up of warm surface waters in the Caribbean region was diminished. However, the southeasterly trade winds and the South Equatorial Current (SEC) are robust features of the climate system (Philander and Pacanowski, 1986) and probably controlled mixed layer depth near Ceara Rise, as in the present day ocean (Fig. 1). Southeasterly trade wind strength is greatest during the late boreal summer when the intertropical convergence zone (ITCZ) is at its northernmost extent, leading to a thick mixed layer and increased strength of the SEC off the northeast coast of South America (Philander and Pacanowski, 1986). Conversely, decreased trade wind strength during boreal spring when the ITCZ is further south leads to a shallower thermocline and reduced SEC strength (Fig. 1). Therefore the Ceara Rise region should be quite sensitive to mixed layer variability related to strength of the SEC and associated southeasterly trade winds.

Correspondingly, the nutrient delivery hypothesis predicts that $\delta^{13}$C maxima should be associated with increased carbonate and/or organic carbon productivity (Raymo, 1994). If increased nutrient delivery came from increased global weathering of continental rocks and sediments, it may be associated with increased bioproductivity across the low-latitude regions. Alternatively, it is possible that increased bioproductivity was limited to coastal and upwelling regions. In either case, increased continental weathering should also have been accompanied by increased radiogenic contributions (e.g., $^{87}$Sr/$^{86}$Sr and $^{187}$Os/$^{186}$Os; Richter *et al.*, 1992; Ravizza, 1993).

![Fig. 1. Location map showing position of Site 926 on Ceara Rise in the western equatorial Atlantic (3°43.148'N, 42°54.507W; 3598.3 mwd). Also shown are average boreal summer position and relative strength of the Southeast Equatorial Current (SEC), North Brazil Coastal Current (NBCC), and North Equatorial Counter Current (NECC).](image-url)
Foraminifera as recorders of mixed layer hydrography and productivity

The relative abundance of mixed layer species in a foraminiferal assemblage has been widely used as a proxy for mixed layer thickness in the western equatorial Atlantic (e.g., Ravel
e et al., 1990; Cullen and Curry, 1997; Chaissen and Ravelo, 1997). Mixed layer species thrive at the expense of thermocline species when the mixed layer is deep, because of increased niche availability in the photic zone. When the thermocline shoals into the photic zone, the relative abundance of thermocline species increases (Leckie, 1939).

The stable isotope geochemistry of planktic foraminiferal tests is also an excellent recorder of the environments in which foraminifera live. Surface-dwelling forms typically exhibit low $\delta^{18}O$ values reflecting warm SSTs and high $\delta^{13}C$ values reflecting nutrient-depleted surface water DIC (e.g., Fairbanks et al., 1982, Ravelo et al., 1990; Ravelo and Fairbanks, 1992; 1995). Thermocline-dwelling forms exhibit higher $\delta^{18}O$ values reflecting colder temperatures and lower $\delta^{13}C$ values reflecting remineralization of organic matter. Differences in $\delta^{18}O$ and $\delta^{13}C$ from coeval samples can be considered to reflect temperature gradients and nutrient gradients from mixed layer to thermocline depths (e.g., Broecker and Peng, 1982). In the modern tropical Atlantic, Globigerinoideas sacculifer (without sac) is commonly used as a recorder of near-surface conditions, and Menardella tumida is often used to record lower mixed layer conditions.

Planktic foraminiferal $\delta^{18}O$ is also influenced by evaporation/precipitation balance. However, salinity variations are expected to be small in this open equatorial ocean setting. This is because the slope of the $\delta^{18}O$-seawater/S relationship is quite low in the equatorial Atlantic (Fairbanks et al., 1992). Secondary calcification at depth is an acknowledged complication (Lohmann, 1995). This along with partial shell recrystallization during early burial may also alter primary isotope values. However, while diageneric influences shift the absolute values, because the majority of primary calcite is retained, the original isotope gradients are largely preserved (Pearson et al., 1993; Crowley and Zachos, 1999; Zachos et al., 2002).

Overall, variability in mixed layer depth and bioproductivity should be manifested in assemblage and isotopic records based on planktic foraminifera that inhabit mixed layer to thermocline depths (e.g., Ravelo et al., 1990; Ravelo and Fairbanks, 1992). In particular, a thick mixed layer that extends below the photic zone should be reflected by (a) dominance of mixed layer species at the expense of thermocline-dwellers, and (b) increased $\delta^{18}O$ gradients between mixed-layer species and thermocline species (Ravelo et al., 1990; Ravelo and Fairbanks, 1992; Pearson et al., 1997; Chaissen and Ravelo, 1997; 2000; Faul et al., 2000). In contrast, a shallower thermocline within the photic zone will (a) reduce the relative abundance of mixed layer foraminifera and (b) decrease the $\delta^{18}O$ gradient between mixed-layer and thermocline species. Similarly, increased productivity in a low-latitude setting should increase the surface-to-thermocline $\delta^{13}C$ gradient as added nutrients enhance the “biologic pump.” Such an increase should be recorded as an increase in the $\delta^{13}C$ gradient between surface and thermocline planktic foraminifera. Increased productivity should also increase abundances of planktic foraminifera and coccolithophorids relative to terrigenous clays in Ceara Rise sediments.

For the O/M boundary interval, we combine full foraminiferal assemblage counts with stable isotope data on Globigerina pataucloides and Paragloborotalia mayeri and investigate their interrelationships with published benthic foraminiferal stable isotope data. These species are chosen because they are interpreted to reflect mixed layer and upper thermocline conditions, respectively (Pearson et al., 1997). Our main goal is to investigate the relation of inferred changes in surface hydrography to global carbon cycling across the O/M boundary.
Fig. 2. Hydrographic profile in the Ceará Rise region from surface to 500 m (after Levitus and Boyer, 1994). Marked by the 18°C isotherm, mixed layer thickness is greatest during the late boreal summer (August shown) in response to increased southeasterly trade winds. Also shown is the δ13C composition of CO₂ at GESECS Stations 39-46 in the western equatorial Atlantic (Kroopnick, 1985), exhibiting high values in surface waters and low values in the thermocline related to bioproducitivity.

Methods

Sediment samples from Hone 926B cores 50X and 51X were taken at 10 cm intervals, washed over a 63 micron sieve, oven-dried at <50°C, and split for assemblage and isotopic work. Full assemblage counts were conducted on quantitative splits of ~300 specimens in the >150 μm size fraction, identified to the species level.

Full planktic foraminiferal assemblage counts were conducted to trace the relative dominance of mixed-layer species vs. thermocline species, which is an index of mixed layer thickness. Mixed layer species are considered to include G. primordius, G. praebulloides, Giobigerinita spp., P. kugleri/pseudokugleri, Tenacellinata angustiumbilicata, and Zeaoglobigerina woodi. Thermocline
species include *Globoquadrina venezuelana*, *G. globularis* and related forms, and *P. mayeri*, while *Catapsydrax dissimilis* is a sub-thermocline species (Pearson et al., 1997; 2001). Species composition is influenced by dissolution, because mixed-layer species are generally more susceptible to dissolution in sediments. Accordingly, the ratio of foraminiferal fragments to fragments plus whole tests was used as an assessment of est breakage, an excellent but labor-intensive dissolution monitor (e.g., Berger et al., 1982; Peterson and Prell, 1985).

*G. praebulloides* and *P. mayeri* from Cores 50X and 51X were analyzed on a Finnegan MAT DeltaPlus XL stable isotope ratio mass spectrometer (SIMS) equipped with a Kiel III automated preparation device in the College of Marine Science at the University of South Florida. External precision based on over 900 NBS-19 standards run on this instrument since July 2000 is better than +/−0.04 ‰ for δ¹³C and +/−0.08 ‰ for δ¹⁸O. Data on the above two species, plus *P. kugleri/pseudokugleri*, *C. dissimilis*, and *G. venezuelana* from Core 50X were generated on a VG Prism S.RMS equipped with an automated carousel preparation device in the Godwin Laboratory at the University of Cambridge.

To investigate the coherence and phasing among our proxy records, Blackman-Tukey cross-spectral analysis was conducted using the Analyseries package (Paillard et al., 1996). We elected not to similarly analyze some of our isotope gradient records, because of a lack of significant orbital-scale variations observed and a poor signal-to-noise ratio in these time series. Each other series was interpolated to a constant 5 k.y. time step, and a cross-correlation function was calculated using 1/3 lags and a Bartlett window with a bandwidth of 6.8 cycles/m.y. This bandwidth encompassed spectral peaks in the eccentricity (125 and 95 k.y.), obliquity (41 k.y.), and precession (23 k.y.) bands.

**Stratigraphy and chronology**

The development of an orbitally tuned chronostratigraphy based on magnetic susceptibility back into the Oligocene (Weedon et al., 1997; Shackleton et al., 1999; Shackleton et al., 2003) allows precise dating in the studied interval (ca. 23.43-22.7 Ma). High-resolution benthic isotope stratigraphy and biostratigraphy have refined the chronology during the late Oligocene to early Miocene (Flower et al., 1997a; Zachos et al., 1997; 2001b), and shown that the O/M boundary coincides with the onset of the M1 glaciation ca. 23.0 Ma (Shackleton et al., 2000). An average sedimentation rate of ~2.6 cm/k.y. and 10 cm sampling allows temporal resolution better than 5 k.y.

**Results and discussion**

**Global carbon cycling and deep-sea temperatures**

Stable isotope data on the planktic foraminifers *G. praebulloides* and *P. mayeri* reveal substantial orbital-scale variability across the O/M boundary. All three species δ¹³C records closely covary during the 22.43-22.7 Ma interval (Fig. 3). Indeed, Pearson et al. (1997) demonstrated that consistent δ¹³C gradients were maintained between five species of planktic foraminifera during the earliest Miocene. Overall, the strong covariance suggests that the δ¹³C records are primarily recording entire water column variations in the δ¹³C composition of DIC, which is best explained by reservoir changes in carbon cycling. In particular, this covariance indicates substantial orbital-scale variability in global organic carbon burial across the O/M boundary. Therefore, we can use any of these records as a first-order proxy for global carbon cycling in this interval. Hodell and Woodruff (1994) also demonstrated that deep sea *Cibicidoides* δ¹³C values were within 0.45 ‰ during the O/M δ¹³C maximum, indicating minimal regional differences due to deep-circulation. In succeeding sections, we use the *Cibicidoides* δ¹³C record as a proxy for global carbon cycling.
Fig. 3. Stable isotope records from Hole 926B vs. age. Planktic foraminiferal $\delta^{13}$C records based on *Globigerina praebulloides* and *Paragloborotalia mayeri* covary with benthic $\delta^{13}$C data based on *Cibicidoides*, reflecting control by global carbon cycling across the O/M boundary. Planktic $\delta^{18}$O records exhibit reduced amplitude relative to *Cibicidoides* $\delta^{18}$O during the M1 $\delta^{18}$O increase, suggesting deep-sea cooling accompanied polar ice growth during M1 (Paul et al., 2000; this study).
Covariance between the oxygen isotope records is weaker, suggesting important deep-sea temperature change during the Mi1 δ¹⁸O increase of ~1‰ at this site (Fig. 3). G. praebulloides δ¹⁸O values range from −2.1 to −0.9 ‰ (mean −1.7 ‰), while P. mayeri values range from −2.1 to −0.6 ‰ (mean −1.4 ‰). Cibicidoides δ¹⁸O exhibits a somewhat greater range (0.7 to 2.0‰), including an interval of distinctly higher values from 23.02-22.9 Ma (Mi1) that represents some combination of deep-sea cooling and Antarctic ice sheet growth (Miller et al., 1991b; Flower et al., 1997a; Zachos et al., 1997; 2001b; Billups and Schrag, 2002). Indeed, the amplitude of Mi1 in the planktic record is half that in the Cibicidoides record, as found previously based on lower resolution data (Paul et al., 2000). This observation may indicate that ~0.5‰ of the Mi1 benthic δ¹⁸O increase was due to deep-sea cooling of ~2°C, or that the ice growth effect in planktic δ¹⁸O was offset by warmer tropical sea surface temperatures (Paul et al., 2000). However, existing benthic Mg/Ca data indicate minimal temperature across the O/M boundary (Billups and Schrag, 2002).

Mixed layer variability

Percent mixed layer species (%MLS) data exhibit significant variations from 10-76%, but do not show a clear relation to Cibicidoides δ¹³C and therefore inferred global carbon burial (Fig. 4). A pair of distinct %MLS increases ca. 23.33-23.26 Ma and ca. 22.94-22.81 Ma do not coincide with Cibicidoides δ¹³C increases. The earlier %MLS increases is associated with a 0.6‰ decrease in Cibicidoides δ¹³C, and the later %MLS increases is associated with the O/M δ¹³C peak. Furthermore, no major increase in %MLS occurs during the initiation of the O/M δ¹³C maximum ca. 23.14 Ma. Similarly, our oxygen isotope proxy for mixed layer thickness reveals no significant change associated with the O/M δ¹³C maximum or Mi1 (Fig. 4). The mixed layer to thermocline δ¹⁸O gradient (G. praebulloides – P. mayeri) shows little mean change over the interval studied, suggesting minimal changes in temperature gradients that might be associated with mixed layer depth.

Significantly, each Cibicidoides δ¹³C maximum (on the 10⁵ year scale) appears to be followed by an increase in %MLS (Fig. 4). Cross-spectral analysis confirms that these records are coherent in the eccentricity band, and that %MLS lags Cibicidoides δ¹³C by 22.6 ± 8.1 k.y. (Table 1). This relationship indicates that mixed layer thickness changed about a quarter cycle (on average) after Cibicidoides δ¹³C. The coherence and consistent phase suggests that trade wind variability was probably not a major driver of global carbon burial in this band, but may have provided a reinforcing feedback through a sustaining influence on coastal upwelling. In contrast, in the precession band %MLS shows a small lead (5.5 ± 3.4 k.y.) vs. Cibicidoides δ¹³C, implying a stronger sensitivity of global carbon cycling to trade wind variability in this band.

Bioproductivity

Our use of Cibicidoides δ¹³C as a monitor of global carbon burial is supported by strong covariance between δ¹³C records from throughout the water column (Pearson et al., 1997; this study). In particular, the mixed layer to deep δ¹³C gradient is maintained; our reconstruction based on G. praebulloides – Cibicidoides δ¹³C records indicates no substantial changes across the O/M boundary (Fig. 5). However, reconstruction of mixed layer to thermocline δ¹³C gradients (G. praebulloides – P. mayeri) shows considerable signal. In particular, this gradient (planktic-planktic δ¹³C in Fig. 5) exhibits seven maxima in ~700 k.y., plus a small increase associated with the initiation of the O/M δ¹³C maximum ca. 23.14 Ma (Fig. 5). Another distinctive feature is a minimum in δ¹³C from ca. 23.35-23.31 Ma that corresponds in part with relatively high Cibicidoides values. We have no ready explanation for this enigmatic event, which is controlled by a marked δ¹³C increase in P. mayeri (Fig. 3). The primary cause of increased mixed layer to thermocline δ¹³C gradients is enhanced extraction of nutrients from surface waters and remineralization at depth (e.g., Broecker and Peng, 1982). An associated increase in bioproductivity can be accomplished with or without addition of nutrients to surface waters by upwelling or riverine delivery.
Fig. 4. Proxies for mixed layer variability and Cibicidoides δ²³C across the O/ M boundary. Neither percent mixed layer species (%MLS) nor the mixed layer: to thermocline δ¹⁸O gradient (G. praebulloides – P. mayeri δ¹⁸O) shows a major increase at the onset of the O/M δ¹³C maximum ca. 23.13 Ma. However, %MLS is coherent with Cibicidoides δ¹³C, and exhibits a small lead in the precession band (Table 1), suggesting modulation of global carbon cycling by trade wind strength in this band.

Cross-spectral analysis confirms coherence between planktic-planktic δ¹³C and Cibicidoides δ¹³C in the eccentricity (100 k.y.) and precessional (23 k.y.) bands (Table 1). Therefore we find some support for the hypothesis that global organic carbon cycling (not necessarily in the western equatorial Atlantic) was accompanied by orbital-scale variations in bioproductivity in this region. Furthermore, a phase lead of 5.8 ± 2.7 k.y. vs. Cibicidoides δ¹³C in the precession band suggest that Ceara Rise bioproductivity changed somewhat earlier (on average) than global carbon cycling in this band. We find no significant coherence in the obliquity band. Overall, our proxies for both mixed layer thickness and bioproductivity are coherent with and lead Cibicidoides δ¹³C in the precession band, consistent with evidence from the late Pleistocene for trade wind influence on equatorial Atlantic Ocean dynamics in this band (McIntyre et al., 1989; Molfin and McIntyre, 1990). Control of trade wind
variability by low-latitude insolation is strong and is related to African monsoon variability (Manabe and Broccoli, 1985; McIntyre et al., 1989; Molfino and McIntyre, 1990). Our results suggest that similar mechanisms were operating across the O/M boundary.

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Blackman-Tukey cross spectral analysis was conducted with the Analyseries program (Paillard et al., 1996) using 1/3 lags and a Bartlett window with a bandwidth of 6.8 cycles/m.y. Coherence is significant at the 80% confidence level (k=0.40). Negative phase (given in degrees and kiloyears; k.y.) indicates the given parameter leads the reference variable.
Fig. 5. Proxies for bioproductivity and Cibicidoides δ¹³C across the O/M boundary. Neither the mixed layer to deep-water δ¹³C gradient (G. praebulloides – Cibicidoides Δδ¹³C) nor the mixed layer to thermocline δ¹³C gradient (G. praebulloides – P. mayeri Δδ¹³C) shows a substantial increase associated with the O/M δ¹³C maximum ca. 23.13 Ma. However, the latter is coherent with Cibicidoides δ¹³C, and exhibits a small lead, in the precession band (Table 1), suggesting modulation of global carbon cycling by bioproductivity in this band.

Nutrient delivery

Previous work has indicated no particular increase in nutrient accumulation in Ceara Rise sediments associated with the O/M δ¹³C maximum (Delany and Anderson, 1997). Increased nutrient supply should be accompanied by increased surface-to-deep water δ¹³C gradients (e.g., Miller and
Fairbanks, 1985; Vincent and Berger, 1985; Delany, 1990), in contrast to our findings. Raymo (1994) suggested a pronounced increase in organic carbon burial during the early to middle Miocene related to Himalayan uplift. Her study found no significant increase near the O/M boundary, but did not attempt to examine organic carbon burial at the orbital scale. Another possible source region is South America. However, Ceara Rise records indicate that sediment delivery by the Amazon was low until ca. 8 Ma, after substantial uplift of the Andes during the middle to late Miocene (Dobson et al., 1997). In contrast, Himalayan uplift and weathering apparently were well underway by the O/M boundary (Uddin and Lundberg, 1993; Najman et al., 1997). Finally, radiogenic isotope proxies for global continental weathering show little increase across the O/M boundary based either on $^{87}$Sr/$^{86}$Sr (Richter et al., 1992; Hodell and Woodruñ, 1994; Oslick et al., 1994) or $^{187}$Os/$^{188}$Os (Ravizza, 1993; Reusch et al., 1998).

On the other hand, our results may be consistent with a corollary of the nutrient delivery hypothesis, that increased nutrient supply during glaciations enhanced surface bioproductivity and thereby the mixed layer to thermocline $\delta^{13}$C gradient in the western equatorial Atlantic. We cannot rule out orbital-scale increases in nutrient delivery controlling bioproductivity. Our suggestion is that nutrient supply (and possibly weathering-induced CO$_2$ drawdown) may have been an important feedback mechanism on the orbital scale for O/M organic carbon burial. An associated decrease in temperature during the Mi1 glaciation may have provided a negative feedback on weathering that led to minimal changes in radiogenic isotope proxies for global continental weathering (Berner et al., 1983). This hypothesis can be tested further by examining orbital-scale variability in bioproductivity throughout Himalayan weathering history. Similarly, the relationships between western tropical hydrography and global carbon burial can be compared in other intervals of enhanced global carbon burial such as the early to middle Miocene. Exploring how orbital-scale changes in wind-driven upwelling and productivity are manifested in non-upwelling areas is a fruitful avenue for research aimed at understanding organic carbon burial.

**Dissolution and deep-water variability**

To monitor the potential effect of seconary processes on our foraminiferal assemblage and stable isotope proxies, we compare proxies for lithologic variability and carbonate dissolution (Fig. 6). Controlled by the relative contribution of carbonate sediments and terrigenous clays, lithologic cycles in cores 50X and 51X (based on %reflectance in the 550 nm band; Fig. 6) are dominated by cyclicity in the 4-k.y. band (Shipboard Scientific Party, 1995). Based on coherent cyclicity in Ba/Al, P/Al, Si/Al, and %CaCO$_3$, Weedon and Shackleton (1997) attributed these cycles to a combination of (1) carbonate bioproductivity, (2) carbonate dissolution, and (3) terrigenous supply. Percent coarse fraction data also indicate a strong influence of dissolution on Ceara Rise sediments, especially in the obliquity band (Flower et al., 1997a; Zachos et al., 1997, 2001; Paul et al., 2000). Fig. 6 shows coherent variations with percent reflectance. However, both percent reflectance and percent coarse fraction data are also influenced by carbonate bioproductivity and terrigenous supply.

Foraminiferal fragmentation provides an independent tracer of carbonate dissolution that has advantages over percent reflectance and percent coarse fraction records employed in earlier studies (Flower et al., 1997a; Zachos et al., 1997, 2001; Pearson et al., 1997; Paul et al., 2000). Our foraminiferal fragmentation proxy (fragments/fragments plus whole planktic foraminifers) overcomes these limitations because it should be influenced only by dissolution. In particular, this proxy is largely controlled by the CO$_2$ concentration in bottom waters (e.g., Broecker and Peng, 1982). Carbonate-corrosive water bathing Ceara Rise near the O/M boundary is inferred to be Southern Ocean Water (SOW) (Wright et al., 1992; Flower et al., 1997b; Paul et al., 2000).
Fig. 6. Proxies for lithologic variability (percent reflectance in the 550 nm band), percent coarse fraction (>63 μm), dissolution (percent planktic fragments/fragments plus whole foraminifera) shown vs. *Cibicidoides* δ¹³C and δ¹⁸O. A long-term increase in percent coarse fraction and percent fragments after 23.1 Ma suggests enhanced carbonate preservation. All three proxies are coherent and in phase in the obliquity band (~41 k.y.), indicating a common control by dissolution (Table 1). However, percent fragments and percent reflectance lag *Cibicidoides* δ¹³C, suggesting global carbon cycling was decoupled from deep-water circulation in this band.
Our fragmentation proxy (Fig. 6) exhibits trends that are independent of both the %MLS and the planktic-planktic Δδ13C records. This proxy is not coherent with either record in any orbital band. This implies that our mixed layer thickness and bioproducitivity records are not fundamentally controlled by dissolution. Pearson et al. (1997) presented evidence for changes in carbonate bioproducitivity based on a Δδ13C proxy similar to that used here, and suggested that its variability was not associated with lithologic cycles. These authors found consistency in the planktic isotope data they generated across lithologic cycles interpreted to reflect dissolution variability. These findings are consistent with the suggestion that although dissolution is significant, it does not appreciably affect either planktic isotope data (Pearson et al., 1997; this study) or the assemblage data (this study).

The percent fragments proxy is coherent and in phase with percent reflectance and percent coarse fraction the obliquity band (Table 1), suggesting a common control by carbonate dissolution. However, percent fragments and percent reflectance also exhibit a significant lag relative to Cibicidoides δ13C of -15-21 k.y. This consistent lag underscores that factors other than SOW strongly influenced Cibicidoides δ13C in this interval, because SOW variability should be accompanied by coeval dissolution changes. Indeed, this relationship is consistent with a phase sequence in which global carbon cycling leads deep circulation changes and dissolution cycles in the deep Atlantic Ocean, at least in the obliquity band.

Finally, comparison of our dissolution index with benthic δ13C and δ18O reveals important long-term trends (Fig. 6). In particular, a transition from high to low percent fragments from ca. 23.36 to 22.9 Ma indicates decreasing dissolution across the O/M δ13C maximum. This finding may turn indicate that the long-term increase in percent coarse fraction beginning during Mi1 was also caused by enhanced preservation more than either increased carbonate productivity or reduced terrigenous input. Furthermore, a distinct dissolution pulse associated with Mi1 at ca. 23 Ma suggests that the initiation of Mi1 was marked by increased SOW at Ceara Rise.

Overall, these results suggest a dominant control of lithologic cycles by carbonate dissolution, tied to SOW variability driven by high latitude forcing. However, Cibicidoides δ13C appears to be decoupled from SOW variability, at least in the obliquity band. Indeed, if Cibicidoides δ13C primarily reflects global carbon cycling, then the carbon cycle must have a complex relationship to Atlantic deep circulation across the O/M boundary.

Conclusions

In Oligocene/Miocene upwelling-dominated settings, it is difficult to ascertain whether organic carbon burial increased because of increased trade wind strength and associated upwelling, increased nutrient supply from continents and associated bioproducitivity, or some other mechanism. Therefore it is advantageous to examine limited- or non-upwelling settings that are sensitive to these processes. Here we attempt to reconstruct western tropical Atlantic hydrography across the O/M boundary to assess the relative importance of trade wind strength and nutrient supply in controlling global organic carbon burial.

Covariance between planktic and benthic δ13C indicates that global carbon cycling controlled water column δ13C in this interval, which can be approximated by Cibicidoides δ13C. Foraminiferal assemblage and isotopic data indicate limited variation in mixed layer thickness and surface bioproducitivity associated with δ13C maxima that comprise the Oligocene/Miocene δ14C excursion. However, in the precession band both percent mixed layer species and the surface to thermocline δ13C gradient are coherent with Cibicidoides δ13C, and exhibit a small lead of ~6 ± 3 k.y. These results are consistent with evidence from the late Pleistocene for trade wind influence on equatorial Atlantic Ocean dynamics in this band (McIntyre et al., 1989; Molino and McIntyre, 1990).
Use of the percent mixed layer species (%MLS) proxy rests on the assumption that mixed layer species did not change their depth preferences due to evolutionary processes in this interval. Testing this assumption is beyond the scope of this paper. However, we speculate that episodic expansion and contraction of the mixed layer may have contributed to the diversification of mixed layer species known from the late Oligocene to early Miocene (e.g., Cifelli, 1969; Kennett and Srinivasan, 1982; Srinivasar and Kennett, 1983; Spezzaferri, 1995).

In the obliquity band, coherence among lithologic variability (based on percent reflectance in the 550 nm band), percent coarse fraction, and a dissolution index (fragments/fragments plus whole planktic foraminifers) indicates a common control by carbonate dissolution. However, none of these is coherent with our surface water proxies, suggesting minimal overprinting by dissolution. Furthermore, percent fragments and percent reflectance exhibit a significant lag relative to Cibicidoides \(^{813}C\) of \(-15\) to 21 k.y. Because Southern Ocean Water (SOW) variability should be accompanied by coeval dissolution changes, this observation implies that other factors strongly influenced Cibicidoides \(^{813}C\) across the O/M boundary. We speculate that global carbon cycling led deep circulation changes and dissolution cycles in the deep Atlantic Ocean, at least in the obliquity band. Our results suggest that SOW variability was decoupled from Cibicidoides \(^{813}C\), and underscores the importance of global carbon cycling in Oligocene / Miocene climate change.

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