Insights from the Paleogene tropical Pacific: Foraminiferal stable isotope and elemental results from Site 1209, Shatsky Rise

Andrea Dutton
Department of Geological Sciences, University of Michigan, Ann Arbor, Michigan, USA

Research School of Earth Sciences, Australian National University, Canberra, ACT, Australia

Kyger C Lohmann
Department of Geological Sciences, University of Michigan, Ann Arbor, Michigan, USA

R. Mark Leckie
Department of Geosciences, University of Massachusetts, Amherst, Massachusetts, USA

Received 14 October 2004; revised 17 April 2005; accepted 22 April 2005; published 28 July 2005.

[1] Stable isotope and elemental measurements were conducted on foraminifera from a sequence of calcareous pelagic ooze at Ocean Drilling Program Site 1209 to document the thermal evolution of the North Pacific water column over Shatsky Rise and to address long-standing questions about the nature of oceanic circulation in the early Paleogene. A major change in seasonality and water column structure in the early Eocene marks a change in tropical Pacific climate and circulation just prior to the early Eocene climatic optimum. Subsequent long-term cooling is interrupted by a transient positive δ18O excursion identified in planktonic and benthic foraminifera during the late early Eocene that is interpreted as evidence for ephemeral formation of deep water from a low-latitude, saline source. Paired analysis of benthic Mg/Ca and δ18O was undertaken to assess the relative contribution of temperature and seawater oxygen isotope composition to the benthic δ18O record. Stratigraphic trends of benthic δ18O and Mg/Ca decouple during two intervals in the early Eocene and early middle Eocene. Although variable seawater δ18O is the most likely candidate to explain decoupling of benthic δ18O and Mg/Ca, it is difficult to argue for substantial change in continental ice volume during a presumably ice-free interval of time.


1. Introduction

[2] In the closing years of the Ocean Drilling Program (ODP), several locations were targeted in the Pacific and Atlantic Ocean basins (ODP Legs 198, 199, 207, and 208) to provide more complete records of extreme climates in the Cretaceous and Paleogene [Lyle et al., 2002; Bralower et al., 2003; Erbacher et al., 2004; Zachos et al., 2004]. The Paleogene is of particular interest to paleoclimatologists because it represents a transition between extreme climate states: from the “greenhouse” climate of the early Eocene climatic optimum (EEOC) to the “icehouse” of the early Oligocene. The equable, warm climates of the early Paleogene were punctuated by at least one transient hyperthermal event at the Paleocene/Eocene boundary [e.g., Kennett and Stott, 1991; Thomas and Shackleton, 1996; Zachos et al., 2003] and the occurrence of multiple hyperthermals in the early Eocene has been further postulated [Thomas et al., 2001]. Thus the early Paleogene contains examples of extreme climates on both short (10⁴ years) and long (10⁶ years) timescales.

[3] Despite the identification of this climatically dynamic period of earth’s history, there are still several questions that persist about the driving mechanisms and feedbacks of Paleogene climate as well as the effects of climatic perturbations upon the biosphere. Paramount among these is the question of how low equator-to-pole temperature gradients can be maintained in the early Paleogene [Barron, 1987; Sloan et al., 1995], particularly in light of evidence that oceanic and atmospheric circulation was sluggish, and therefore less capable of transporting heat to polar regions [Janecek and Rea, 1983; Rea, 1998]. A warm sluggish ocean with low vertical thermal gradients and no winter sea ice to facilitate downwelling of deep waters at high latitudes may have been prone to circulation reversals whereby warm, salty water sinks at lower latitudes. The viability of this hypothesis continues to be heavily debated in the literature [Bice et al., 1997; Bice and Marotzke, 2001 and references therein], nearly a century after it was proposed by Chamberlain [1906]. Finally, the driving mechanisms of both long- and short-term climate changes in the Paleogene are still questioned, as well as the interaction and feedbacks between processes that operate on different timescales. Our collective ability to address these questions has been hampered in part by the coarseness of spatial and temporal...

Copyright 2005 by the American Geophysical Union. 0883-8305/05/2004PA001098
sampling of well-preserved Paleogene deep-sea sedimentary sections. To address this deficiency, this investigation was designed to provide a multiproxy documentation of long-term climatic and oceanographic history from a tropical site in the Pacific Ocean. The Pacific Ocean basin was much larger in the Paleogene (Figure 1), yet because of the relatively shallow depth of the carbonate compensation depth (CCD), particularly in the northern Pacific, recovery of well-preserved carbonate sediment from this region is rare.

ODP Site 1209 (32°39.1081′N, 158°30.3564′E) was drilled on the Southern High of Shatsky Rise in a water depth of 2387 m. The sedimentary record recovered at Site 1209 contains an expanded Paleogene section consisting of unlithified calcareous ooze with moderately to well-preserved foraminifera. This site represents the shallowest in a depth transect of sites targeting Paleogene sediments cored during Leg 198, and was situated well above the CCD during most of the Paleocene and Eocene [Hancock and Dickens, 2005]. Stable isotope analyses were carried out on several planktonic taxa (Acarinina, Morozovella, Globigerinatheka, Praemurica, Subbotina) as well as several benthic taxa (Nuttalides, Oridorsalis, Cibicidoides, Gavelinella, Lenticulina). A record of Mg/Ca for benthic foraminifera has also been constructed to better evaluate the relative influence of temperature and oxygen isotope seawater composition upon the benthic oxygen isotope record.

2. Methods

2.1. Sample Preparation

The samples studied are from the composite section of Site 1209, which splices together sections from three holes (A, B, and C) [Bralower et al., 2003]. Approximately one 10-cm³ sample per section was selected for analysis; these were dried overnight in an oven at <50°C, disaggregated in buffered deionized water (pH ~10), washed through 63- and 125-μm sieves, and oven dried again. The >125μm fraction was dry sieved into four fractions: >125–250 μm, >250–300 μm, >300–355 μm, and >355 μm. Planktonic foraminifera were picked from a narrow size fraction (300–355 μm) to minimize complicating factors of ontogenetic and other vital effects upon geochemical compositions [Shackleton et al., 1985; Pearson et al., 1993]. In seven samples planktonics were taken from the 250–300 μm size fraction. Benthic foraminifera were picked from the >125 μm size fraction.

2.2. Foraminiferal Taxa

Species of Acarinina and Morozovella were selectively chosen from the surface-dwelling taxa present because they are abundant, morphologically distinctive, and frequently analyzed from other Paleogene deep-sea cores. In two samples, these taxa were not present and Praemurica inconstans and Globigerinatheka sp. were used instead. Because no single species of Acarinina or Morozovella spans the entire time interval studied, several different species were analyzed, including: Morozovella conicotruncana, M. angulata, M. subbotinae (including M. gracilis), M. velascoensis, M. formosa, M. aragonensis, Acarinina soldadoensis, and A. bulbrooki. Where possible, multiple taxa were examined for cross-calibration purposes. Species of Subbotina were grouped together for analysis, although most Subbotina samples consist of a single species. Each stable isotope analysis consists of four to seven individual planktonic foraminifera.

Several benthic taxa were used for stable isotope analyses, including: Gavelinella spp., Nuttalides spp., Oridorsalis spp., Lenticulina spp., and Cibicidoides spp. Of these taxa, only Nuttalides truempyi, Oridorsalis umbonatus, and Gavelinella beccariformis were used for elemental analyses. Multiple taxa were analyzed in several cases for both types of analyses to determine offsets in geochemical composition between species. Depending upon size and abundance, between one and five benthic specimens were used for stable isotope analysis and between 11 and 28 individual benthics were combined for elemental analysis.
Table 1. Relevant Geographic Information of Deep-Sea Sites Mentioned in the Text

<table>
<thead>
<tr>
<th>Site</th>
<th>Location Name</th>
<th>Present Depth, m</th>
<th>Paleodepth, m</th>
<th>Paleodepth Zone</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Paleolatitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>ODP 1209</td>
<td>Shatsky Rise</td>
<td>2187</td>
<td>2500(^{a})</td>
<td>upper abyssal</td>
<td>33°N</td>
<td>159°E</td>
<td>15°–20°N</td>
</tr>
<tr>
<td>DSDP 577</td>
<td>Shatsky Rise</td>
<td>2675</td>
<td>2400(^{b})</td>
<td>upper abyssal</td>
<td>32°N</td>
<td>158°E</td>
<td>15°–20°N</td>
</tr>
<tr>
<td>ODP 865</td>
<td>Allison Guyot</td>
<td>1518</td>
<td>1300–1500(^{c})</td>
<td>lower bathyal</td>
<td>18°N</td>
<td>180°E</td>
<td>2°–6°N</td>
</tr>
<tr>
<td>ODP 689</td>
<td>Maud Rise</td>
<td>2080</td>
<td>1400–1650(^{d})</td>
<td>lower bathyal</td>
<td>65°S</td>
<td>3°E</td>
<td>67°S</td>
</tr>
<tr>
<td>ODP 757</td>
<td>Ninetyeast Ridge</td>
<td>1652</td>
<td>1500(^{f})</td>
<td>lower bathyal</td>
<td>37°S</td>
<td>88°E</td>
<td>45°S</td>
</tr>
<tr>
<td>DSDP 550</td>
<td>Goban Spur</td>
<td>4432</td>
<td>4000(^{f})</td>
<td>lower abyssal</td>
<td>48°N</td>
<td>13°W</td>
<td>48°N</td>
</tr>
</tbody>
</table>

\(^{a}\)Bralower et al. [2003];
\(^{b}\)Zachos and Arthur [1986];
\(^{c}\)Thomas and Shackleton [1996];
\(^{d}\)Kennett and Stott [1990];
\(^{e}\)Pierce et al. [1989];
\(^{f}\)Masson et al. [1985].
**3. Results**

**3.1. Stable Isotope Data**

**3.1.1. Surface-Dwelling Planktonics**

[13] Oxygen isotope compositions of surface dwellers, species of *Acarinina* and *Morozovella*, maintain a relatively stable long-term stratigraphic trend centered on -1.4‰ throughout the Paleocene, but show substantial variability within this interval (Figure 2). Single specimen isotopic analyses of eleven *M. velascoensis* specimens from an upper Paleocene sample (224.5 mcd) display a range in compositions that is roughly an order of magnitude greater than the analytical precision of each analysis, spanning 0.5‰ in δ¹⁸O and 0.9‰ in δ¹³C. The double-headed arrows in Figure 2 depict this range of intrasample variability, which is similar in magnitude to the variability between stratigraphically adjacent samples. Low δ¹⁸O values reaching -1.7‰ are recorded by lowermost Eocene *Acarinina* and *Morozovella*. These low values are followed by a positive shift of ~+0.3‰ (199.3 mcd). From this point in the lower Eocene, δ¹⁸O gradually increases into the upper middle Eocene where a maximum value of ~0.2‰ is attained. The variability between stratigraphically adjacent samples is typically much smaller through the Eocene than in the Paleocene.

**3.1.2. Deeper-Dwelling Planktonics**

[14] Carbon isotope compositions of surface-dwelling planktonics steadily increase from the lower Paleocene to the upper Paleocene where they peak at maximum values of 4.9 to 5.1‰. After reaching this apex, δ¹³C rapidly decreases by ~3.5‰ through the uppermost Paleocene and lowermost Eocene. δ¹³C values stabilize in the lower Eocene and display gradual enrichment in ¹³C from lower to middle Eocene, increasing from roughly 2.0 to 2.7‰.

---

*Figure 2.* Foraminiferal (left) oxygen isotope and (right) carbon isotope data plotted relative to mean composite depth (mcd) for Site 1209. The double-headed arrows at ~225 mcd indicate the range in composition of 11 single specimen analyses from the same sample.

---

*F. tympaniformis* set at 55.0 Ma by Bralower [2005] was excluded. Magnetostratigraphic control for this sequence is not available.
upper Paleocene, reaching a minimum of −1.3‰ in the lowermost Eocene (208.8 mcd). This minimum is immediately followed by a 1.0‰ increase to higher δ18O values that prevail for most of the lower Eocene. A return to low δ18O values occurs in one Subbotina sample (183.6 mcd) near the lower to middle Eocene boundary, but a dearth of well-preserved Subbotina immediately above this sample precludes us from establishing whether low values are maintained or if this one sample is anomalously low relative to its stratigraphic neighbors. The middle Eocene exhibits high δ18O values from −0.4 to +1.2‰ that generally increase upward section with the exception of a notable excursion to lower values in the upper middle Eocene (160.1 mcd).

[16] Carbon isotope compositions of Subbotina display oscillatory values in the Paleocene, and steadily increase from 1.7‰ in the lower Paleocene to 3.3‰ in the upper Paleocene. Maximum values are attained during the same stratigraphic interval that records the highest values for the surface-dwelling planktonics. These high δ13C values decay rapidly through the uppermost Paleocene and lowermost Eocene, reaching a minimum of 0.6‰ that is coincident with the minimum attained in the surface planktonic δ13C record (205.2 mcd). After reaching this minimum, Subbotina displays gradually increasing δ13C values during the lower Eocene that lead into a sharp increase of 0.8‰ (193.0 mcd). After this point, Subbotina δ13C values remain steady at about 1.7‰ from the lower into the middle Eocene.

3.1.3. Benthic Foraminifera

[17] Benthic oxygen isotope values shown in Figure 2 have been adjusted for interspecies offsets. We used the corrections calculated for Nuttalides, Cibicidoides, and Oridorisalis by Katz et al. [2003] who carried out a rigorous assessment of intertaxa offsets for Paleogene specimens. Corrections for Nuttalides and Cibicidoides have been applied to this data set as well as to all benthic oxygen isotope data from other deep-sea cores cited herein. We chose to use these corrections because they yield better coherency between values recorded by multiple taxa from the same sample and/or within the same stratigraphic interval both in our data set and for Paleogene data in the global compilation of Zachos et al. [2001]. Similar adjustments were made for the carbon isotope data using offsets from Katz et al. [2003], in part because of the combination of infaunal (Oridorisalis, Lenticulina) and epifaunal taxa (Nuttalides, Cibicidoides, Gavelinella) in this study. These taxa display significant offsets in δ13C because the infaunal taxa are exposed to 13C-enriched pore waters. We note that species-adjusted δ13C also reduces the apparent scatter in the Site 1209 data as well as in the Zachos et al. [2001] Paleogene data set. Lenticulina and Gavelinella were not included in the Katz et al. [2003] study; therefore additive correction factors were calculated for these taxa based on average offsets observed in the Shatsky Rise data set (δ18O, +0.3‰ for both species; δ13C, +1.24‰ for Lenticulina and +0.36‰ for Gavelinella).

[18] Paleocene benthic foraminifera display gradual oscillations in δ18O that range from 0.1 to 0.8‰. A rapid decrease occurs in δ18O values from about 0.5 to −0.4‰ in the uppermost Paleocene and lowermost Eocene (213.3 to 206.7 mcd). Benthic δ18O stabilizes at low values between −0.4 and 0.0‰ for the duration of the lower Eocene and the lower middle Eocene with the exception of one sample (1209B-20H-4, 51–53 cm; 193.0 mcd), represented by three different benthic species, that records exceptionally high δ18O values of 0.7 to 0.8‰. This positive δ18O excursion of ~1.0‰ is accompanied by positive excursions in Acarinina and Subbotina that are both 0.5‰ in magnitude. Low δ18O values that characterize most of the lower and lower middle Eocene rapidly increase from −0.4 to 0.9‰ in the middle Eocene (176.0 to 162.4 mcd). A second anomalous sample (1209C-6H-4, 40–42 cm; 160.1 mcd) is observed in the middle Eocene which displays a negative δ18O excursion of ~1.1‰, and again is accompanied by an excursion in Acarinina, Morozovella, and Subbotina. The surface dwellers only decrease by ~0.5‰ in this horizon while Subbotina δ18O diminishes by 0.8‰.

[19] Trends in carbon isotope values of benthic foraminifera at Shatsky Rise are remarkably similar to trends recorded in Subbotina δ13C. Benthic δ13C increases from the lower to upper Paleocene attaining a maximum value of 2.1‰ that is largely coincident with peak values recorded in planktonic foraminifera. These high δ13C values rapidly decrease across the Paleocene/Eocene boundary and reach minimum values in the lowermost Eocene. The δ13C then gradually recovers to higher values through the lower Eocene and is marked by a distinctive positive step of 0.7‰ in the upper lower Eocene, similar in timing and magnitude to the stepwise increase observed in Subbotina specimens. The δ13C slowly decreases back to values averaging 0.5‰ that remain relatively constant through the remainder of the middle Eocene.

3.2. Elemental Data

[20] Mg/Ca values for all taxa were normalized to Oridorisalis umbonatus by determining the difference between pairs of species for several samples. Additive corrections using fixed values cannot be used to normalize Mg/Ca data in the way that benthic foraminiferal δ18O data is typically treated because the relationship between Mg/Ca and temperature is exponential. This temperature dependence can be described in a general manner by the following equation:

\[ Mg/Ca = B \exp(A \times T) \]

where Mg/Ca is in mmol/mol, T is in °C, A is the exponential constant that represents temperature sensitivity and B is the preexponential constant. Thus, the difference in Mg/Ca between two species is a function of the temperature of calcification (Figure 3). As such, it is imperative to derive an exponential function for the relation between Mg/Ca and temperature for each extinct species. This procedure is illustrated for Nuttalides truempyi in Figure 3. In this diagram, the temperature for seven paired analyses of N. truempyi and O. umbonatus Mg/Ca measurements was determined using the Mg/Ca-temperature equation for O. umbonatus from the core top calibration of Lear et al. [2002]. Additional paired analyses of these two species from the study of Lear et al. [2000] are also included. This exercise demonstrates that a number of exponential functions encompassing the range of temperature sensitiv-
In all diagrams displaying Mg/Ca data, species-corrected Mg/Ca data have been converted to temperature using the core top calibration for *O. umbonatus* [Lear et al., 2002] to facilitate comparison with oxygen isotope data that are linearly related to temperature. Absolute temperatures calculated from Mg/Ca data depend on the assumed Mg/Ca composition of Paleogene seawater, yet relative change in temperature is insensitive to this parameter. Mg/Ca displays variability that corresponds to ~2°C in temperature during the Paleocene, although the long-term trend is relatively constant during this interval and in the early Eocene (Figure 4). Mg/Ca values, and their corresponding inferred temperatures, decrease gradually from late early to middle Eocene. This cooling trend begins at ~50 Ma and continues through the remainder of the section studied (to ~40 Ma). Fluctuations in benthic Mg/Ca mimic patterns observed in benthic δ¹⁸O throughout much of the section except for the earliest Eocene and the early middle Eocene where Mg/Ca and δ¹⁸O decouple. These intervals are both marked by relatively rapid change in δ¹³C relative to the Mg/Ca record.

4. Discussion

4.1. Foraminiferal Preservation

Paleocene and Eocene foraminifera at Site 1209 are generally moderately to well preserved with increasing effects of dissolution upward toward the upper Eocene [Hancock and Dickens, 2005; Petrizzo et al., 2005]. A minor component of secondary calcite was evident in some SEM images [Dutton et al., 2005]. The volume of recrystallized calcite in the best preserved specimens is less than 5% and ranges up to ~10% for the more poorly preserved specimens. Individuals that were strongly recrystallized and samples dominated by fragments and corroded calcite were not selected for geochemical analysis, including all samples from the upper Eocene. Objective evidence such as maintenance of intertaxa offsets and similarity to other deep-sea data sets combined with a more subjective visual assessment of shell textures suggests that the geochemical signatures of these foraminifera are largely unaffected by diagenesis. Nonetheless, diagenesis can be subtle and can elude standard procedures aimed at elucidating its effects. Therefore we cannot completely rule out the possibility that post depositional alteration has had some effect on the geochemical data generated from Site 1209.

4.2. Intertaxa Offsets in Stable Isotope Composition

24] Systematic differences between δ¹³C and δ¹⁸O of planktonic taxa can indicate the influence of different depth habitats, vital effects, or season of calcification. *Acarinina* and *Morozovella* are both considered to have lived in the surface mixed layer [e.g., Pearson et al., 1993] but their relative depth habitats are debatable. Offsets reported for *Acarinina* and *Morozovella* from several different Paleogene deep-sea cores are inconsistent from site to site and sometimes show different patterns through time [Oberhänsli and Toumarkine, 1985; D’Hondt et al., 1994; Bralower et al., 1995a]. Paired measurements of these taxa at other sites
do not always display systematic offsets [Pearson et al., 1993]. At Site 1209, early Eocene A. soldaoensis does not show systematic differences in comparison to M. subbotinae. Early to middle Eocene A. bulbrooki is 0.1% lower in d18O than M. aragonensis in four different samples, but no consistent offset is observed in d13C. 

The relation between these surface dwellers and deeper dweller Subbotina is much clearer. Subbotina consistently records more positive d18O values and more negative d13C than any of the surface-dwelling taxa throughout the section at Site 1209. This systematic pattern of offsets in both isotope records has been inferred as an indicator of a depth habitat within or below the thermocline [Stott et al., 1990; Pearson et al., 1993; D’Hondt et al., 1994]. Interestingly, Subbotina records from Shatsky Rise are similar to those at Allison Guyot [Bralower et al., 1995b] in that more short-term variability is observed for this taxon than for the other planktonic taxa. This variability could be attributed to mobility in depth habitat, temporal fluctuations in environmental conditions of deep surface waters, or may result from the combination of multiple species used to generate this record. Of these possibilities, we suggest that fluctuating temperatures in thermocline waters is the most likely explanation. This idea is consistent with predictions of a coupled ocean atmosphere climate model, which describes an El Niño–like oscillation in the tropical and subtropical Pacific during the early Paleogene [Huber, 2002].

4.2. Benthic Foraminifera

We measured stable isotope compositions of both Nuttalides and Oridorsalis in eight samples. Removing one outlier from this group, we observe an excellent correlation between their d18O values (r² = 0.97): d18O_NUTT = (d18O_ORID - 0.44)/0.89. This relation is very similar to that determined by Katz et al. [2003] which is derived from a larger data set compiled from several cores. Carbon isotope values of epifaunal benthic taxa (Cibicidoides, Gavelinella, Nuttalides) are more positive than Oridorsalis d13C values in 15 out of 17 cases. This result agrees with the findings of other studies that suggest a shallow infaunal habitat for Oridorsalis [e.g., Rathburn and Corliss, 1994]. Although we have applied the correction of Katz et al. [2003] to normalize d13C of Oridorsalis to Cibicidoides, we note that this produces values that are somewhat high relative to the other taxa throughout the section (Figure 2). This result indicates that the d13C adjustment for Oridorsalis may be slightly overcorrection d13C values relative to compositions that are in equilibrium with dissolved organic carbon of ambient seawater.

4.3. Stable Isotope Paleoclimate Reconstruction

During the Paleocene and Eocene, trends in planktonic d18O at Site 1209 are largely similar to those at other sites [Boersma et al., 1987; Stott et al., 1990; Barrera and Huber, 1991, 1993; Zachos et al., 1994; Bralower et al., 1995b]. Surface water foraminifera record low but relatively stable d18O values during the Paleocene and earliest Eocene that correspond to a temperature of ~ 22°C (Figure 2). Considerable intrasample and intersample variability in d18O during this time interval is interpreted as the result of a seasonal climate. This pattern is contrasted by low intersample variability during the late early and middle Eocene that may reflect a substantial decrease in the seasonal amplitude of SST. Planktonic oxygen isotope...
values steadily increase from the early Eocene through the middle Eocene, signifying the initiation of a long-term cooling trend. Temperatures drop from 22°C to 17°C in surface waters from ~52 to 45 Ma (202.2 mcd to 156.6 mcd). This surface cooling trend precedes cooling in the benthic record, which attains minimum δ18O values (maximum warmth) at roughly 52 Ma yet does not begin to increase (cool) until after 48 Ma (176.1 mcd). The significance of anomalously negative planktonic and benthic δ18O values at 45.4 Ma (160.1 mcd) is not clear, but may correlate to a transient negative excursion seen in the planktonic isotope record from Site 865 in the equatorial Pacific [Bralower et al., 1995b].

There is a notable association of strong surface-to-benthic δ18O gradients over Shatsky Rise with more positive values of benthic δ18O (Figure 5). The collapse of the δ18O gradient in the early Eocene is accomplished primarily by a substantial decrease in benthic δ18O and is accompanied by a positive shift in Subbotina δ18O (Figure 2). The shift observed in Subbotina is similar in timing, magnitude, and direction to the shift in Subbotina at DSDP Site 550 in the abyssal North Atlantic [Charisi and Schmitz, 1996]. The near elimination of a gradient between thermocline and deep waters seen at Site 1209 (~53 Ma) is also recorded at Site 550. This reduction in the vertical δ18O gradient precedes initiation of long-term cooling of both surface (after 52 Ma) and deep waters (after 48 Ma), but is coincident with minimum δ13C values recorded by benthic and planktonic foraminifera at Site 1209 and is also coincident with the change from high seasonality in the surface waters to low seasonality (as demonstrated by the intersample variability of Acarinina and Morozovella values). This point (52.7 Ma, 206.7 mcd) marks the termination of a long-term decrease in benthic δ13C and δ18O in the late Paleocene and earliest Eocene and may signify a change in ocean circulation at this time. The surface-to-thermocline (Subbotina) δ18O gradient remains relatively stable in comparison to the surface-to-benthic gradient. A strong vertical gradient in δ18O is reestablished as benthic δ18O begins to increase in the upper middle Eocene.

Surface-to-benthic carbon isotope gradients reach a maximum in the late Paleocene, similar to observations at other deep-sea sites [Corfield and Cartlidge, 1992]. The vertical δ13C gradient reaches a minimum in the late early Eocene but increases thereafter. In this case, the surface-to-thermocline and surface-to-benthic gradients parallel each other throughout the section (Figure 5). Differences be-

![Figure 5. Vertical oxygen and carbon isotope gradients in the water column over Shatsky Rise plotted relative to age for Site 1209. Gradients were calculated using average values of surface planktonics, deeper dwelling planktonics (Subbotina), and benthics for each sample.](image-url)
tween surface and benthic $\delta^{13}$C result from the preferential uptake of $^{12}$C by photosynthesis in the surface ocean, thus changes in the magnitude of this gradient should be a reflection of the amount of organic carbon exported to the deep ocean. If the partitioning of $^{12}$C between surface and deep water over Shatsky Rise is interpreted in this way, we infer peak productivity in the late Paleocene is associated with a thermally well-stratified but highly seasonal upper water column, while minimum productivity is recorded in the early Eocene near the apex of $\delta^{18}$O (warmth) in the benthic record and reduced seasonality in the surface record.

There are a number of isotopic studies of Paleogene foraminifera from DSDP Site 577 on Shatsky Rise [Shackleton et al., 1985; Zachos and Arthur, 1986; Miller et al., 1987; Corfield and Cartlidge, 1992; Pak and Miller, 1992; Lu and Keller, 1996]. The Paleogene section recovered at ODP Site 1209 is more complete than the section recovered at Site 577, thus a major goal of this study was to temporally extend this isotopic record and to pair benthic $\delta^{18}$O data with Mg/Ca to evaluate the relative influence of temperature and water composition on the oxygen isotope record. Where our data set overlaps with temporally equivalent data from Site 577, agreement between absolute values and trends of $\delta^{18}$O and $\delta^{13}$C is excellent with the exception of the Subbotina data. Site 1209 Subbotina isotopic values agree well with those of Shackleton et al. [1985], but these data are offset from compositions reported by Corfield and Cartlidge [1992]. To some extent this discrepancy may be attributed to differences between taxa, but because all of these studies used multiple species of Subbotina this is unlikely to account for systematic offset between these different data sets. Differences between size fractions cannot explain this observation as Shackleton et al. [1985] found no significant ontogenetic trend in stable isotope compositions of Subbotina. The more likely, yet unsettling interpretation is that Subbotina measurements on samples taken several years after core collection have experienced alteration. Subbotina may be more susceptible to alteration after core recovery than other planktonic or benthic taxa because of its coarsely perforate shell. We emphasize that this interpretation is tentative but may call for further investigation into this type of phenomenon.

4.4. Evidence for Saline Deep Water

Evaluation of deep water isotopic signals is an important step toward constraining the evolution of circulation patterns and thermal history of the Paleogene. Modern ocean circulation is driven by thermohaline processes whereby dense surface waters in the Southern Ocean and the North Atlantic sink to form deep water. The globally warm climate of the early Cenozoic may have further affected ocean circulation patterns by lowering the density of high-latitude surface waters and causing low-latitudinal gradients in SST. This may have allowed for the production of warm, saline bottom water (WSBW) derived from low latitudes [Chamberlain, 1906; Brass et al., 1982; Kennett and Stott, 1990]. Though many investigators have sought direct evidence for such a reversal in ocean circulation, indisputable evidence of such an event has remained elusive. In addition to differences in thermal gradients, the
connectivity of ocean basins in the Paleogene was also quite different than the modern configuration. Continental rearrangements during this time frame affected several oceanic gateways, including the opening of the northern North Atlantic, Tasman Sea, and the Drake Passage, as well as the closing of the Tethys [Scotese et al., 1988]. These changes may well be associated with concomitant change in climatic regime and ocean circulation.

[33] To place the results of our study in the context of global climate and oceanic circulation, we first compare Site 1209 benthic isotope data to the compilation of Zachos et al. [2001] (Figure 6). For comparative purposes, isotopic values from these records have been normalized as described in section 3.1.3. Long-term trends in both δ18O and δ13C at Site 1209 are quite similar to those at other sites, but some differences exist.

[34] In the modern ocean, the oldest deep waters are in the North Pacific Ocean where 14C-enriched waters result from the contribution of increasing amounts of oxidized organic matter on the seaﬂoor. This results in a δ13C gradient between the Paciﬁc Ocean and the Southern, Atlantic, and Indian Ocean basins that have younger deep waters. During the early Paleogene, carbon isotope values of Paciﬁc sites, including Site 1209, are generally more negative than those from other ocean basins except during a brief interval of the Eocene from ~50 to 48 Ma when the interbasinal gradient is minimal and all ocean basins display a positive shift in δ13C (Figure 6). It has been argued that ampliﬁed basinal gradients in δ13C during the Paleocene and Eocene reﬂect a stronger inﬂuence of Southern Ocean−derived deep water whereas diminished basinal gradients result from possible downwelling of warm saline bottom water derived from the Tethys or the tropical/subtropical Paciﬁc [Corﬁeld and Cartlidge, 1992; Pak and Miller, 1992]. This interpretation would lead us to speculate that during the interval from ~50 to 48 Ma there may be a component of WSBW. This notion is substantiated by a fully coupled model of early Eocene (~50 Ma) climate that indicated the presence of relatively warm, salty deep water that formed primarily in the Tethys and the North Atlantic Oceans [Huber and Sloan, 2001].

[35] At 50.5 Ma, an anomalously positive δ18O value is recorded in foraminifera at all depths in the water column over Shatsky Rise (Figures 2 and 7). Both surface and deeper dwelling planktonics increase in δ18O by 0.5‰ whereas all three benthic taxa analyzed increase by roughly 1‰. Although only one sample shows this shift, we believe that it represents a significant change in ocean circulation caused by a rapid switch to more saline deep water that resulted in high values of δ18O. Although the planktonic foraminifera for this sample show slightly more secondary calcite than samples taken from the sections immediately above and below, the benthics appear well preserved under binocular and scanning electron microscope examination. If the aberrant values were entirely a product of diageneis, mixing between primary calcite and a positive diageneric end-member should induce more of a shift in the composition of the planktonic taxa, yet the positive shift observed is actually larger in magnitude for the benthics. We therefore interpret that the benthic taxa grew in equilibrium with saline bottom waters and that the planktonic taxa either (1) grew in somewhat saline surface waters that were less saline than the source of deep water or, (2) having settled out on the sediment-water interface, were subject to early diagenesis that produced secondary calcite and shifted their overall δ18O to higher values.

[36] Additional evidence for production of deep water from a saline source in the early Eocene is provided by comparison of data from Site 1209 in the subtropical Paciﬁc and Site 865, which was located in the equatorial Paciﬁc during the Paleogene (Figure 7). The inferred switch in circulation is associated with a global shift toward more positive δ13C as well as a collapse in the vertical carbon isotope gradient at Site 865 (Figures 6 and 7). These patterns suggest that 13C-enriched surface water in the tropical Paciﬁc may have been the source of saline deep water. This downwelling may have had more impact on intermediate water formation than on true deep-water formation depending on the depth to which these waters penetrated. Our sampling resolution constrains this downwelling to less than 0.5 m.y. in duration. Interestingly, during this time interval surface planktonic foraminifera at Site 865 record high temperatures and high salinities on the basis of a Mg/Ca−based temperature reconstruction [Tripatie et al., 2003] (Figure 7). These Mg/Ca data require a positive shift of ~1.0‰ in surface water δ18O over Allison Guyot because of salinity; this is the same amount of change in benthic δ18O at Shatsky Rise that we have attributed to salinity. The coincidence of a brief hiatus at Site 865 [Bralower and Mutterlose, 1995] during this circulation reversal event lends further support for a change in deep-water currents at this time, although the question of why saline deep waters should form in the tropical Paciﬁc rather than in the subtropical belt is a question that needs to be addressed.

[37] Model-predicted locations of deep water convection at ~50 Ma appear to be controlled by salinity gradients suggesting a circulation pattern that is “quasi-halothermal” [Huber and Sloan, 2001]. Such an ocean circulation regime would be susceptible to alteration by sudden changes net evaporative patterns and would be expected to produce deep water masses that are more salty, but not necessarily colder than overlying intermediate water. Another model of the early Paleogene ocean predicts that subduction of low-latitude, saline surface waters would have been ephemeral, and may result in the delivery of saline intermediate waters to high latitudes that could lead to a reintensification of high-latitude deep water formation [Bice and Marotzke, 2001]. This progression from weak low-latitude sinking of saline water to intensification of convection at middle and high latitudes is referred to as “flushing” [Weaver and Sarachik, 1991] and is predicted to lead to cooling of the deep ocean. It is possible that a brief period of subduction of saline surface water in the late early Eocene led to deep ocean cooling; this hypothesis is supported by benthic Mg/Ca and δ18O records at Site 1209 which both show the initiation of cooling following the inferred switch in circulation (Figure 4).
Instability of the water column at low latitudes during the early Paleogene may have resulted in episodic downwelling that is difficult to resolve without intense sampling resolution. If WSBW penetrated the depths of the Paleogene ocean, we would expect to see this effect upon a number of chemical, sedimentary and biological systems. In fact, the sulfur isotope record from marine barite displays a dramatic shift which begins at the Paleocene-Eocene boundary, but accelerates rapidly at 50 Ma [Paytan et al., 1998]. This shift is interpreted by Kurtz et al. [2003] to represent a peak in pyrite burial flux at 50 Ma as a consequence of warmer source waters for the deep ocean. Evidence for warm, euxinic deep water at this time is further supported by Kaiho [1991], who calculated an oxygen index through faunal analysis of benthic foraminifera and found that oxygenation of the deep ocean drops to a minimum at the Paleocene/Eocene boundary and remains low until the benthic δ18O signal rapidly increases at 50 Ma. Both of these records indicate that warm water began to penetrate to depth in the oceans at the Paleocene-Eocene boundary, and that an abrupt return to more oxygenated deep water occurs near the end of the early Eocene. These data complement our hypothesis of an oceanic circulation reversal in the late early Eocene and suggest that weak, downward advection of warm water, possibly only to intermediate water depths, began in the early Eocene. The strength in our hypothesis of WSBW formation during the EECO lies in the temporal alignment of perturbations in several proxies from multiple sites including carbon isotopes, oxygen isotopes, Mg/Ca ratios, sulfur isotopes, and a faunal proxy for deep-water oxygenation. We expect that future high-resolution studies during this time interval should help to clarify the interpretation put forth here.

4.5. Early Eocene Hyperthermal Events

The Paleocene-Eocene thermal maximum at Shatsky Rise is recognized by a shoaling of the CCD and lysocline along the depth transect of cores recovered, and it is characterized by a dramatic drop in carbonate content, a temperature spike of 5°C in surface waters, and a sea surface salinity anomaly of +0.5‰ [Bralower et al., 2003; Zachos et al., 2003; Colosimo et al., 2005]. Although the existence of multiple such hyperthermal events in the early Paleogene has been postulated [Thomas et al., 2001], direct evidence for this hypothesis was lacking until multiple conspicuous clay layers associated with peaks in the magnetic susceptibility record were identified at both Shatsky Rise and Walvis Ridge [Bralower et al., 2003; Zachos et al., 2004]. We observe a very negative benthic δ13C value at 52.7 Ma (206.7 mcd), which may be correlative to the “Chron 24n event” clay layer observed at Walvis Ridge [Zachos et al., 2004] (Figure 7). Further high-resolution study of these hyperthermal events should help to better constrain their timing, duration, and magnitude.

4.6. Benthic Mg/Ca Record at Shatsky Rise

To this point, we have made preliminary inferences of δ18O data assuming that this proxy primarily records temperature of the ambient water. However, it is well known that foraminiferal oxygen isotope data are subject to the influence of temperature as well as the oxygen isotope...
the Mg/Ca ratio of seawater (Mg/CaSW) may be variable. Peng 
Mg/CaSW values of 3 to 4 mol/mol for this period of time. 
Intervals are characterized by rapid change in 
d Regardless of the absolute value of Mg/CaSW that we apply, 
the relative influence of temperature and water composition 
sensitive to change in continental ice volume. To investigate 
d Mg/CaSW of 3.3 mol/mol. This estimate is in agreement 
globally, measured values of Mg/Ca at Site 1209 imply a 
benthic data is debatable, the Mg/Ca data are shown relative to the 
summary given by Wilkinson and Algeo [1989] model, which predicts 
Mg/Ca values of echinoderm calcite [Dickson, 2002], and by two 
models [Wilkinson and Algeo, 1989; Stanley and Hardie, 1998]. The error associated with each of these techniques is 
quite large and therefore uncertainty in the Mg/Ca of Paleogene seawater is significant (see summary given by 
Billups and Schrag [2003]). Nonetheless, all of these attempts to reconstruct the evolution of Mg/CaSW agree that 
Mg/Ca of seawater was lower in the Paleogene. 
As the appropriate value of Mg/CaSW to apply to our 
data is debatable, the Mg/Ca data are shown relative to the 
benthic δ18O record using two estimates of seawater composition (Figure 4). Assuming that the world was ice-free at 
∼50 Ma, when minimum benthic δ18O values are recorded globally, measured values of Mg/Ca at Site 1209 imply a 
Mg/CaSW of 3.3 mol/mol. This estimate is in agreement with the Wilkinson and Algeo [1989] model, which predicts 
Mg/CaSW values of 3 to 4 mol/mol for this period of time. Regardless of the absolute value of Mg/CaSW that we apply, 
our data display a distinct decoupling between Mg/Ca and 
δ18O during two intervals in the early (∼55–53 Ma) and early middle Eocene (∼48–45 Ma). Both of these time intervals are characterized by rapid change in δ18O and relatively little change in Mg/Ca. The long-term trend in benthic water temperature according to Mg/Ca-derived estimates is quite different than that for δ18O-derived paleotemperatures. The Mg/Ca record indicates little change in average temperature from ∼63 to 50 Ma, and displays a gradual decline in temperature beginning at ∼50 Ma. This long-term trend may be punctuated by short-term pulses in temperature that are not captured by this relatively coarse resolution record. 
If we assume that Mg/Ca is a reliable paleothermometer then we must conclude that the classic trough in benthic δ18O of the early Eocene that is taken to represent peak Cenozoic warmth (i.e., the EECO) is not a product of increased warmth and requires another explanation. This idea seems unlikely at first glance for several reasons, the most apparent of which is that the most likely candidate for change in δ18O apart from temperature is ice volume, and these data would require melting and subsequent accumulation of ∼50–80% of modern ice volume in a climate considered to be entirely ice-free! Given these considerations, we present several scenarios that could explain the observation of decoupling between Mg/Ca and δ18O of benthic foraminifera from Shatsky Rise. 
4.6.1. Hypothesis 1: Diagenesis 
Diagenesis of carbonate sediment in the deep-sea occurs as dissolution of primary calcite and/or precipitation of secondary calcite (also termed recrystallization). Dissolution has the effect of lowering the Mg/Ca ratio of foraminiferal calcite [Brown and Elderfield, 1996; Rosenthal et al., 2000] whereas recrystallization should increase the Mg/Ca ratio because the distribution coefficient of Mg in abiotic calcite is greater than that of biotic calcite [e.g., Mucci and Morse, 1983]. In either case, diagenesis should induce convergence of Mg/Ca toward a diagenetic calcite end-member. Yet, Mg/Ca and δ18O vary sympathetically through much of the section, with the exception of the two intervals of decoupling described earlier (Figure 4). This indicates that relative temperature changes were responsible for the patterns observed and that these signals have not been diagenetically homogenized. Moreover, Mg/Ca and δ18O are altered proportionately during dissolution of foraminiferal calcite, which preserves their relationship to temperature [Rosenthal et al., 2000]. Hence paired measurements of Mg/Ca and δ18O can still be used to reconstruct δ18O even in partially dissolved specimens but even more importantly, dissolution is not a mechanism that can explain decoupling of these two proxies. 
4.6.2. Hypothesis 2: Variable Seawater Mg/Ca 
Our initial comparison of relative change in Mg/Ca and δ18O at Site 1209 was made with the tacit assumption that Mg/CaSW was invariant. Yet, the residence times of Mg and particularly Ca in the ocean require that we entertain the possibility that secular variation in Mg/CaSW can explain decoupling of the benthic Mg/Ca and δ18O records. This argument has several weaknesses: (1) The data would require extremely rapid (1–2 m.y.) changes in seawater composition separated by a protracted period of invariant Mg/CaSW and these rapid changes must be coupled with rapid temperature change to produce foraminiferal calcite that has approximately the same value before and after both of these rapid changes. (2) Amplified hydrothermal activity in the early Eocene [Leinen, 1987] would have the effect of lowering Mg/CaSW to produce the first decoupling, but would require a large change in Mg/CaSW (from 5.2 to 3.3 mol/mol) and available data on seafloor spreading rates imply declining rates of hydrothermal activity at this time [e.g., Kominz, 1984; Gaffin, 1987]. (3) The trend in benthic Mg/Ca parallels that of planktonic δ18O and records the same temperature as subthermoline waters for Mg/CaSW = 3.3 mol/mol (Figure 8). This conformity of long-term temperature trends at several depths in the water column suggests that the benthic Mg/Ca temperature reconstruction is genuine, and unaffected by substantial change in Mg/CaSW during this interval. This observation is particularly interesting and turns our attention to the fidelity of the benthic oxygen isotope record as a proxy for temperature. 
4.6.3. Hypothesis 3: Ice Volume 
It seems likely that perturbations to the δ18O signal should explain the decoupling observed between Mg/Ca
The proposed change in pH is well beyond the range of concentrations proposed for the last glacial maximum. Figure 8. Reconstructed paleotemperatures for benthic foraminifera (solid circles), deep-dwelling planktonics (i.e., *Subbotina*, squares), and surface-dwelling planktonics (open circles). Benthic paleotemperatures are calculated using Mg/Ca compositions; planktonic paleotemperatures are calculated using δ¹⁸O data and δw = -0.98‰ (Subbotina) and -0.14‰ (surface dwellers).

and δ¹⁸O because benthic δ¹⁸O changes rapidly during these intervals while Mg/Ca remains relatively stable through the Paleocene and changes gradually through part of the Eocene record. Although the decay and subsequent regrowth of a large ice sheet in the Eocene is the mostly likely mechanism to drive the change in benthic δ¹⁸O, such a large volume of ice (50–80% of modern ice volume) in the globally warm early Paleogene world is not very plausible. Not only is there a lack of geological and paleontological evidence for an ice sheet of this size in the early Paleogene, climate studies from this time interval indicate a very warm, equable climate characterized by low gradients in pole-to-equator surface temperatures [e.g., Zachos et al., 1994]. Unless direct evidence of such ice is discovered we must acknowledge that the scenario for significant ice volume in the late Paleocene and early middle Eocene is difficult to prove independently.

4.6.4. Hypothesis 4: pH

[47] In addition to temperature and seawater oxygen isotope composition, the δ¹⁸O proxy is also sensitive to pH [Spero et al., 1997]. To induce a negative shift of 0.8‰ in δ¹⁸O would require an increase in carbonate ion concentration, [CO₃²⁻], of 400 mmol/kg and a concomitant decrease in δ¹³C of 2.4‰ according to the modern calibration of planktonic foraminifera [Spero et al., 1997]. Although δ¹³C does decrease by roughly this amount, the proposed change in pH is well beyond the range of [CO₃²⁻] in the modern ocean as well as the range proposed for the last glacial maximum [Broecker and Peng, 1982; Sanyal et al., 1995]. Moreover, there is not evidence for a significant deepening of the carbonate compensation depth in the early Eocene Pacific basin, though there are few records that span this time period [Van Andel, 1975; Lyle et al., 2002; Hancock and Dickens, 2005]. One further weakness of this hypothesis is that change in pH has been demonstrated to decrease planktonic Mg/Ca by 6% per 0.1 pH unit [Lea et al., 1999]. If a similar relationship exists for benthic Mg/Ca and pH, then variable pH in the deep ocean cannot explain decoupling of Mg/Ca and δ¹³C.

4.6.5. Hypothesis 5: Opening of an Oceanic Gateway

[48] We submit one final hypothesis for a mechanism capable of altering δw by 0.8‰ in the early Paleogene that calls upon the dynamic rearrangement of the global ocean during this time interval. The opening of several major oceanic gateways mentioned earlier might have established a deep-water connection to a large, restricted basin of significantly different oxygen isotope composition. Such a scenario could account for an influx of water depleted in ¹⁸O that would be capable of affecting the benthic δ¹³C signal in all ocean basins. This explanation is appealing because the opening northern North Atlantic would have established a connection to the Arctic Ocean basin, which likely would have had a much lower δw, due in part to the entrainment of ¹⁸O-depleted, high-latitude meteoric water. Whether there was a large enough volume of water and whether it was depleted enough to induce a large negative shift in benthic δ¹³C of all ocean basins is debatable.

4.6.6. Comparison to Other Paleogene Mg/Ca Records

[49] To this point, we have focused on comparison of Mg/Ca and δ¹³C of benthic foraminifera at Site 1209 without consideration for other benthic Mg/Ca records from this time interval. Only two other studies have been conducted representing two somewhat shallower sites in the Southern Ocean (ODP Site 689) and the Indian Ocean (ODP Site 757) (Table 1 and Figure 1) [Lear et al., 2000; Billups and Schrag, 2003]. For purposes of comparison, data from these studies have been converted to temperature using modern core top calibrations of Lear et al. [2002] and assuming a modern Mg/CaSW of 5.2 mol/mol (Figure 9).

[50] If we assume that ODP Sites 689, 757, and 1209 all represent an intermediate to deep water signal then their benthic Mg/Ca records should be comparable. Site 1209 shows excellent agreement with the Lear et al. [2000] Mg/Ca data where these records overlap. In contrast, both records generated by Billups and Schrag [2003] display dramatic increase in Mg/Ca (and hence in reconstructed temperatures) down core. This is somewhat paradoxical because the Lear et al. [2000] data during this time interval is entirely based upon ODP Site 689, the same site investigated by Billups and Schrag [2003]. The difference in these records is presumably due to (1) potential differences in cleaning protocols; (2) accuracy of interlaboratory Mg/Ca data; (3) unrecognized size and/or ontogenetic differences in Mg/Ca of benthic foraminifera; or (4) inappropriate application of modern core top calibrations to extinct taxa. Unfortunately, it is difficult to discuss comparisons between
the Site 1209 benthic Mg/Ca record and these other two studies until an adequate explanation can be provided for the differences between two records generated from the same sediment core.

5. Summary and Conclusions

Foraminiferal geochemistry results from Site 1209 highlight a number of critical intervals of change in oceanic circulation and climate during the Paleocene and Eocene. An important positive $\delta^{18}O$ excursion (50.5 Ma) has been identified during the EECO that we interpret to represent an ephemeral switch in oceanic circulation patterns to a deep-water source derived from saline surface waters in the tropical Pacific. The data presented here also highlight the transition at 52.7 Ma as an important change from high- to low-temperature seasonality in surface waters, coincident with a collapse in the thermocline-to-benthic $\delta^{18}O$ gradient that persisted until ~45 Ma, and marked by an extremely negative benthic $\delta^{13}C$ value that we tentatively correlate to the “Chron 24n event” clay layer observed at Walvis Ridge [Zachos et al., 2004]. This juncture also represents the attainment of the Cenozoic minimum in benthic $\delta^{18}O$ referred to as the EECO. Increasing Subbotina $\delta^{18}O$ values at Sites 1209, 865, and 550 may be an indication that saline waters began to penetrate thermocline depths at this time, which preconditioned the ocean for an abrupt reversal in oceanic circulation at ~50.5 Ma.

The most important lesson from this investigation is that this multiproxy reconstruction of deep ocean temperatures in the early Paleogene provides two very different scenarios of the thermal history of the ocean. The generation of additional benthic Mg/Ca records from other sites will undoubtedly help to support or refute the interpretations put forth here with respect to the thermal evolution of deep-water masses during the early Paleogene. Intriguingly, benthic Mg/Ca from ODP Site 1209 on Shatsky Rise predicts upper abyssal water temperatures of $8^\circ$–$10^\circ$C (using modern Mg/CaSW) in the early Eocene which is a good match to temperatures predicted by climate models [Huber, 2002]. While this agreement between models and data does not necessarily validate the Mg/Ca proxy, this comparison highlights the critical need to generate multiproxy records for Paleogene climate in order to better interpret the $\delta^{18}O$ signal and to discern key modes of oceanic circulation of this ancient greenhouse climate.

Acknowledgments. We gratefully acknowledge T. Bralower, S. Schellenberg, and L. Sloan for constructive comments; L. Vermuelen, L. Wingate, and L. McMorrow for assistance with sample preparation and analysis; C. Lear for providing her Mg/Ca data; and M. Greaves for supplying aliquots of elemental standards. This research used samples provided by the Ocean Drilling Program (ODP). ODP is sponsored by the U.S. National Science Foundation (NSF) and participating countries under management of Joint Oceanographic Institutions (JOI), Inc. Funding for this research was provided by a JOI-USSAC award and a Rackham Predoctoral Fellowship to A.D.
Mucci, A., and J. Morse (1983), The incorporation of Mg$^+$ and Sr$^+$ into calcite overgrowths: Influences of growth rate and solution com-
Oberhänsli, H., and M. Tourkamink (1985), The Paleogene oxygen and carbon isotope history of Sites 522, 523, and 524 from the central South Atlantic, in South Atlantic Paleoceneo-
Pak, D. K., and K. G. Miller (1992), Paleocean to Eocene benthic foraminiferal isotope as-
semblages: Implications for deepwater circula-
Pearson, P. N., N. J. Shackleton, and M. A. Hall (1993), Stable isotope paleoecology of middle Eocene planktonic foraminifera and multi-
Pierc, J. W., et al. (1989), Proceedings of the Ocean Drilling Program Scientific Results, Leg 121, Ocean Drill. Program, College Sta-
tion, Tex.
Rathburn, A. E., and B. H. Corliss (1994), The ecology of living (stained) deep-sea benthic foraminifera from the Sula Sea, Paleocean-
ography, 9, 87–150.
Rea, D. K. (1998), Changes in atmospheric circulation during the latest Paleocene and earliest Eocene epochs and some implica-
Scotese, C. R., L. M. Gaahagan, and R. L. Larson (1988), Plate tectonic reconstructions of the Cretaceous and Cenozoic ocean basins, Tecto-
nopgraphy, 155, 27–48.
Shackleton, N. J., R. M. Corfield, and M. A. Hall (1985), Stable isotope data and the ontogeny of Paleocene planktonic foraminifera, J. For-
Stanley, S. M., and L. A. Hardie (1998), Secular oscillations in the carbonate mineralogy or reef-building and sediment producing organ-
isms driven by tectonically forced shifts in sea-
water chemistry, Palaeogeogr. Palaeoclimatol. Palaeoecol., 144, 3–19.
Stott, L. D., J. P. Kennett, N. J. Shackleton, and R. M. Corfield (1990), The evolution of Ant-
artic surface waters during the Paleogene: In-
Thomas, E., and N. J. Shackleton (1996), The Paleocene-Eocene benthic foraminiferal ex-
tinction and stable isotope anomalies, in Cor-

Zachos, J. C., et al. (2004), Proceedings of the Ocean Drilling Program Initial Reports, vol. 208, Ocean Drill. Program, College Sta-
tion, Tex.

DUTTON ET AL.: SITE 1209—STABLE ISOTOPES AND Mg/Ca