New evidence for subtropical warming during the late Paleocene thermal maximum: Stable isotopes from Deep Sea Drilling Project Site 527, Walvis Ridge

Deborah J. Thomas and Timothy J. Bralower

Department of Geological Sciences, University of North Carolina, Chapel Hill

James C. Zachos

Department of Earth Sciences, University of California, Santa Cruz

Abstract. The late Paleocene thermal maximum (LPTM) was a dramatic, short-term global warming event that occurred ~55 Ma. Warming of high-latitude surface waters and global deep waters during the LPTM has been well documented; however, current data suggest that subtropical and tropical sea surface temperatures (SSTs) did not change during the event. Conventional paradigms of global climate change, such as CO₂-induced greenhouse warming, predict greater warming in the high latitudes than in the tropics or subtropics but, nonetheless, cannot account for the stable tropical/subtropical SSTs. We measured the stable isotope values of well-preserved late Paleocene to early Eocene planktonic foraminifera from South Atlantic Deep Sea Drilling Project (DSDP) Site 527 to evaluate the subtropical response to the climatic and environmental changes of the LPTM. Planktonic foraminiferal $\delta^{18}\text{O}$ values at Site 527 decrease by ~0.94 ‰ from pre-LPTM to excursion values, providing the first evidence for subtropical warming during the LPTM. We estimate that subtropical South Atlantic SSTs warmed by at least ~1° - 4°C, on the basis of possible changes in evaporation and precipitation. The new evidence for subtropical SST warming supports a greenhouse mechanism for global warming involving elevated atmospheric CO₂ levels.

1. Introduction

One of the most dramatic global warming events in the geologic record occurred near the end of the Paleocene epoch (~55 Ma). This event, the late Paleocene thermal maximum (LPTM) [Zachos *et al.*, 1993], was characterized by profound changes in the Earth's climate, oceans, and biosphere. During the LPTM, Antarctic sea surface temperatures (SSTs) warmed from ~14° to 20°C, and deep-ocean temperatures increased from ~11° to 17°C [Stott *et al.*, 1990; Kennett and Stott, 1991]. The deep-sea warming led to the largest extinction of benthic foraminifera of the past 90 million years [e.g., Tjalsma and Lohmann, 1983; Thomas, 1990]. While benthic foraminifera experienced extinctions, other faunal groups diversified, including the planktonic foraminifera [Kelly *et al.*, 1996, 1998] and terrestrial mammals [e.g., Wing *et al.*, 1991; Hooker, 1996]. Marine and terrestrial carbon isotope records demonstrate a 3 - 4 ‰ negative excursion coinciding with the LPTM [e.g., Kennett and Stott, 1991; Koch *et al.*, 1992], reflecting a major change in the nature of global carbon cycling. The dramatic nature of the LPTM is enhanced by the fact that all of these environmental changes occurred within a time span of 50 - 100 kyr [e.g., Thomas and Shackleton, 1996].

There are only a few complete and unaltered deep-sea sedimentary records of the LPTM, most of which lie within temperate to high latitudes. High-latitude LPTM warming is well

documented [e.g., Kennett and Stott, 1991]; however, reconstruction of the tropical and subtropical responses to the event has been hindered by a lack of geographic coverage, incomplete recovery (e.g., Deep Sea Drilling Project (DSDP) Site 577, Shatsky Rise, Pacific Ocean [Aubry, 1995]), and foraminiferal diagenesis (e.g., Ocean Drilling Program (ODP) Sites 999 and 1001, Caribbean Sea [Bralower *et al.*, 1997]). ODP Site 865 in the equatorial Pacific is the only tropical LPTM record with well-preserved planktonic foraminifera, but the oxygen isotope record suggests that SSTs did not change across the LPTM [Bralower *et al.*, 1995]. Several expanded LPTM land sections from the subtropical Tethyan region have been investigated, but foraminifera from these sections typically suffer from diagenetic alteration [e.g., Charisi and Schmitz, 1995; Schmitz *et al.*, 1997], precluding the interpretation of oxygen isotope records in a paleoclimatic or paleoceanographic context.

There are few plausible causal mechanisms for the dramatic and rapid environmental changes that occurred during the LPTM. Conventional mechanisms fail to explain the combination of global warming (concentrated at temperate to high latitudes), the perturbation of the carbon cycle, and the diverse biotic responses. Warming of the deep oceans may have resulted from a switch in deep water sources from cool high-latitude regions to warmer and more saline lower-latitude areas [Kennett and Stott, 1990]. Warmer, less oxygenated deep waters could have caused the benthic foraminiferal extinction (BFE) [e.g., Kennett and Stott, 1991; Thomas and Shackleton, 1996].

The rapid deep water warming is believed to have caused the thermal dissociation of methane hydrates, releasing an es-

timated $1.4 - 2.8 \times 10^{18}$ g of isotopically light ($\sim 60\%$) carbon into the water column [e.g., *Dickens et al.*, 1995, 1997]. Methane hydrate dissociation is the only feasible mechanism able to explain the magnitude and abrupt onset of the carbon isotope excursion (CIE). In addition, the release of methane may have fueled greenhouse warming and contributed to carbonate dissolution and deep water dysoxia [*Dickens et al.*, 1997].

The LPTM is presumed to have resulted from a major reorganization in ocean circulation; however, the ultimate trigger of this reorganization remains elusive. North Atlantic Igneous Province (NAIP) volcanism may have been responsible for long-term global warming that began around 61 Ma [e.g., *Eldholm and Thomas*, 1993]. Enhanced atmospheric CO_2 concentrations resulting from the effusive outgassing of the NAIP may have preferentially warmed the high latitudes, triggering the switch to low-latitude downwelling during the LPTM [*Eldholm and Thomas*, 1993]. Alternatively, the deep-sea circulatory change may have been caused by one or more large-scale, explosive Caribbean volcanic eruptions that could have cooled subtropical and tropical latitudes relative to high-latitude regions [*Bralower et al.*, 1997].

All of the postulated causes of the LPTM remain speculative, largely because of the lack of modern analogs. Massive dissociation of methane hydrates can only be inferred from the CIE. Considerable uncertainty surrounds the timing and climatic effect of NAIP volcanism with respect to the timing of the LPTM [e.g., *Eldholm and Thomas*, 1993; *Thomas and Shackleton*, 1996]. The climatic effect of large-scale, explosive Caribbean volcanic eruptions is also difficult to constrain because their inferred magnitude is far greater than any historical eruption [*Bralower et al.*, 1997]. Perhaps the aspect of the LPTM that has proven most difficult to explain is that the apparent "global warming" event did not seem to affect tropical and subtropical SSTs [e.g., *Stott*, 1992; *Bralower et al.*, 1995]. The lack of evidence for low-latitude sea surface warming is at odds with model simulations of greenhouse warming because enhanced atmospheric CO_2 concentrations should warm the entire globe [e.g., *Sloan and Thomas*, 1999].

An important step toward understanding the cause(s) of the LPTM is the acquisition and evaluation of proxy data from subtropical and tropical sites. Such records will directly address whether the lack of observed tropical and subtropical

Table 1. Location and Paleodepth of the Sites Investigated in This Study

Site	Paleolatitude	Paleodepth, m
527	35°S	3400
690	65°S	1900
865	5°N	1500
999	10°N	1750
1001	10°N	2500

warming was a true characteristic of the LPTM or whether more geographic coverage is required to assess the tropical and subtropical response. Here we present LPTM planktonic stable isotope records from DSDP Site 527 (Walvis Ridge, South Atlantic; paleolatitude $\sim 35^\circ\text{S}$) and apply them toward the evaluation of the subtropical South Atlantic response to the LPTM event. To evaluate the completeness of the LPTM section at Site 527 as well as the effect of diagenetic alteration on the stable isotope records, we also include preservational data and calcium carbonate records from Site 527 as well as from high-latitude ODP Site 690 (Maud Rise, Southern Ocean) and tropical ODP Sites 865 (Allison Guyot, central Pacific), 999 (Kogi Rise, Caribbean Sea), and 1001 (Colombia Basin, Caribbean Sea) (Figure 1 and Table 1).

2. Methods

2.1. Sedimentological and Preservational Studies

Samples were soaked in a 70:30 solution of calgon (2 g L^{-1}) and hydrogen peroxide (10%), placed on a shaker table overnight, and then rinsed over a 63 μm sieve with buffered (pH 10) deionized water. The $>63 \mu\text{m}$ fraction was briefly ultrasonicated in buffered water and then rinsed again. Calcium carbonate contents of bulk samples from Sites 527 and 690 were determined with a Coulometrics coulometer at the University of North Carolina, Chapel Hill (UNC-CH), with a reproducibility of $\pm 0.10\%$.

Preservational analyses from Sites 527, 690, 865, and 1001 (locations in Figure 1 and Table 1) were conducted to constrain potential diagenetic alteration of stable isotope values. While the records from Site 527 are the primary focus of this investigation, preservational data from the other sites were included to provide a basis for comparison as well as to demonstrate the variability in foraminiferal preservation. Specimens of *Acarinina soldadoensis* were broken open and inspected in a Leica S440 scanning electron microscope (SEM) at UNC-CH. In addition, trace element analyses were conducted on samples consisting of 25 foraminiferal specimens from the 250–300 μm size fraction. The trace element samples were ultrasonicated in buffered (pH 10), deionized water to mechanically remove any clay particles from the inner chambers and then rinsed three times in ultrapure deionized water and broken open with a clean wooden pick. Next the samples were reductively cleaned using the technique of *Boyle* [1981]. Cleaned samples were dissolved in 6 mL of 1 N nitric acid. Sr and Ca were analyzed with a Fisons Spectraspan 7 direct current plasma spectrophotometer at UNC-CH using instrument calibration standards prepared from Spex brand pure,

Figure 1. Paleogeographic reconstruction of the late Paleocene [after *Zachos et al.*, 1994] with the locations of the five Deep Sea Drilling Project (DSDP)/Ocean Drilling Program (ODP) sites investigated and/or discussed in the text.

single-element plasma-grade solutions in a nitric acid matrix, yielding a precision of $\pm 2\%$ (2σ).

2.2. Stable Isotope Strategy

Typically, planktonic foraminiferal stable isotope records are constructed from monospecific or monogeneric analyses of the 300–355 μm sieve size fraction in order to constrain ontogenetic effects on stable isotopic composition and allow intersite comparison of isotopic records [Shackleton *et al.*, 1985]. However, the ~ 30 cm LPTM claystone interval at Site 527 contained no specimens larger than 250 μm .

To construct the most complete isotope record possible, three sets of overlapping stable isotope records were generated, each based upon a different size fraction (Figure 2): 300–355 μm (*A. soldadoensis*), 150–250 μm (*Acarinina* spp.), and 90–150 μm (*Acarinina* spp.). Horizons containing *A. soldadoensis* tests from the 300–355 μm fraction extended from 203.30 to 201.10 meters below sea floor (mbsf), and from 200.66 through 191.71 mbsf. Overlapping records from the 150–250 and 90–150 μm size fractions were generated for the intervals from 201.30 through 200.41 mbsf. This interval spans the claystone as well as the underlying and overlying chalk, providing “background” isotopic values for comparison to those derived from within the claystone interval. Severe calcium carbonate dissolution precluded the construction of continuous planktonic foraminiferal stable isotope records, resulting in a 20 cm gap at the onset of the LPTM (Figure 2 and Table 2).

In preparation for stable isotope analyses, specimens were loaded into stainless steel boats and roasted under vacuum at 380°C for 1 hour. Samples were analyzed via an Autocarb device coupled to a Fisons Prism mass spectrometer (University of California, Santa Cruz), in which each sample is reacted in a common acid (phosphoric) bath at 90°C . All values are reported in the delta (per mil) notation relative to the Pee Dee

belemnite (PDB) standard. External analytical precision based on replicate analyses of two standards NBS-19 and Carrera Marble (an in-house standard) was better than 0.08‰ for both O and C isotopes.

2.3. Age Model

Even though Paleogene timescales and absolute ages are in a state of revision [e.g., Aubry *et al.*, 1988; Berggren *et al.*, 1995], a uniform age model was derived for this study to enable correlation of the LPTM geochemical records. Thomas and Shackleton [1996] assigned 55.5 Ma to the BFE and 55.44 Ma to the sedimentary horizon in which the slope of the carbon isotope curve approaches zero. In this study the 1–2 cm stratigraphic interval containing the most negative $\delta^{13}\text{C}$ excursion value was selected as the 55.5 Ma age point because in many sections the maximum $\delta^{13}\text{C}$ excursion value could be more tightly constrained stratigraphically than the BFE. Constant sedimentation rates were assumed between the two LPTM tiepoints (55.5 and 55.44 Ma) as well as between overlying and underlying biostratigraphic and magnetostratigraphic datums (Table 3). Ages for the biostratigraphic and magnetostratigraphic datums are from Berggren *et al.* [1995].

3. Results

3.1. Planktonic Foraminiferal Preservation

Preservational data are presented for Sites 527, 690, 865, and 1001. Representative SEM images of Sites 527 and 1001 foraminifera from above and below the LPTM dissolution intervals [Bralower *et al.*, 1997; this study], as well as Site 690 LPTM interval foraminifera are shown in Figure 3. Specimens from the LPTM claystone interval at Site 527 are shown in Figure 4. Samples from Site 999 are too lithified to isolate individual specimens. Preservational analysis of Site 865 specimens was limited to Sr/Ca measurements because the pristine wall structure has already been documented [Bralower *et al.*, 1995].

Foraminiferal preservation at Site 527 is excellent in the chalk horizons above and below the LPTM clay interval (Figures 3a and 3b). Pore spaces are open (lack diagenetic infill), and there is only minor evidence of calcitic overgrowth on the wall surface. However, planktonic foraminifera from the Site 527 LPTM claystone interval contain varying abundances of zeolite crystals (Figure 4). Despite the presence of zeolites the calcite wall structure is well preserved, with only minimal overgrowth and pore space infilling. Planktonic foraminiferal preservation in the LPTM interval at Site 690 is excellent (Figures 3c and 3d), while Site 1001 foraminifera outside the LPTM interval display more variable and generally much poorer preservation than the specimens from Sites 527 and 690 (Figures 3e and 3f). Extensive overgrowth, pore infilling (Figure 3e), and large rhombs of recrystallized calcite (Figure 3f) are common.

Planktonic foraminiferal Sr/Ca values from Sites 527, 690, 865, and 1001 are shown in Figure 5 and Table 4. Sites 527, 690, and 865 have consistently higher Sr/Ca values than Site 1001. Sr/Ca values of foraminifera from Sites 527, 690, and 865 are between 1 and $1.5 \text{ mmol mol}^{-1}$, and the Site 1001 values are $\sim 0.8 \text{ mmol mol}^{-1}$.

Figure 2. Site 527 CaCO_3 record (right) and the stable isotope strategy employed in this study (left). The bracketed arrows indicate the stratigraphic range of the planktonic foraminiferal stable isotope records.

Table 2. Site 527 Carbonate and Stable Isotope Data

Sample	Depth, mbsf	Age, Ma	CaCO ₃ , %	300-355 μ m		150-250 μ m		90-150 μ m	
				$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
23-2, 55-56 cm	191.55	54.611	82.06						
23-2, 71-72 cm	191.71	54.626	77.44	2.44	-0.95				
23-2, 86-87 cm	191.86	54.641	83.37	2.55	-0.80				
24-1, 27-28 cm	199.27	55.342	86.86	2.84	-0.37				
24-1, 49-50 cm	199.49	55.362	90.43	2.44	-0.38				
24-1, 73-74 cm	199.73	55.385	82.38	1.71	-0.08				
24-1, 96-97 cm	199.96	55.407	87.52	2.46	-0.37				
24-1, 110-111 cm	200.10	55.420	90.64	2.32	-0.30				
24-1, 131-132 cm	200.31	55.440	90.33	2.18	-0.43				
24-1, 141-142 cm	200.41	55.453	82.98	2.13	-0.49	1.85	-0.58	1.44	-0.34
24-2, 4-5 cm	200.54	55.470	85.79	1.72	-0.59	1.49	-0.65	1.19	-0.48
24-2, 9-10 cm	200.59	55.477	78.09	1.80	-0.87	1.36	-0.84	0.88	-0.90
24-2, 12-14 cm	200.62	55.480	62.75	2.00	-1.08	0.99	-1.06	0.81	-1.15
24-2, 16-18 cm	200.66	55.486	51.51	1.49	-1.38	0.81	-1.36	0.46	-1.31
24-2, 21-23 cm	200.71	55.492	51.78			0.91	-1.38	0.27	-1.44
24-2, 25-27 cm	200.75	55.497	50.14			0.78	-1.32		
24-2, 27-28 cm	200.77	55.500				0.44	-1.50	-0.22	-1.78
24-2, 30-32 cm	200.80	55.501	30.80			0.70	-1.49	0.33	-1.58
24-2, 34-36 cm	200.84	55.502	3.73						
24-2, 34-39 cm	200.85	55.503				0.10	-1.54		
24-2, 37-39 cm	200.87	55.504	1.96						
24-2, 43-45 cm	200.93	55.506	1.33						
24-2, 43-56 cm	201.00	55.508						1.89	-0.83
24-2, 51-53 cm	201.01	55.508	0.23						
24-2, 54-56 cm	201.04	55.509	1.19						
24-2, 56-58 cm	201.06	55.510	0.86					1.82	-0.86
24-2, 58-60 cm	201.08	55.511	68.72			1.72	-0.54	1.65	-0.74
24-2, 60-62 cm	201.10	55.512	69.14	3.15	-0.74	2.69	-0.60	1.90	-0.54
24-2, 64-66 cm	201.14	55.513	80.00	3.50	-0.78	2.69	-0.05	2.20	-0.62
24-2, 66-68 cm	201.16	55.514	83.13			2.55	-0.61	2.06	-0.62
24-2, 68-70 cm	201.18	55.514	80.52	3.60	-0.97	2.34	-0.72	2.04	-0.79
24-2, 73-75 cm	201.23	55.516	83.66	3.23	-0.77	2.59	-0.67	1.99	-0.61
24-2, 80-82 cm	201.30	55.519	82.64	3.26	-0.59	2.55	-0.59	2.27	-0.50
24-2, 95-97 cm	201.45	55.524	88.13	3.65	-0.79				
24-2, 106-108 cm	201.56	55.528	86.37	3.57	-0.66				
24-2, 116-118 cm	201.67	55.532	86.84	3.78	-0.74				
24-3, 10-11 cm	202.10	55.547	87.74	3.66	-0.90				
24-3, 41-42 cm	202.41	55.557	78.50	3.73	-0.83				
24-3, 73-74 cm	202.73	55.569	86.92	3.71	-0.88				
24-3, 130-131 cm	203.30	55.589	87.35	3.61	-0.62				

Table 3. Biostratigraphic and Chemostratigraphic Datums Used in the Age Model

Datum	527	690	865	999	1001	Age, Ma
FAD <i>Tribrachiatulus contortus</i>	189.00 ^a					54.37
FAD <i>Tribrachiatulus bramlettei</i>		150.04 ^b				55.00
LAD <i>Fasciculithus tympaniformis</i>			100.90 ^c	971.84 ^d	234.10 ^d	55.33
CIE recovery	200.31	169.40	102.10	975.04	237.76	55.44
Peak of CIE	200.77	170.35	102.90	975.57	239.00	55.50
FAD <i>Discoaster multiradiatus</i>	220.75 ^a	185.49 ^b	114.60 ^c	1000.81 ^d	267.79 ^d	56.20

All stratigraphic positions are in mbsf. Ages from *Berggren et al.* [1995].
 FAD and LAD signify first appearance datum and last appearance datum, respectively.
 CIE refers to the carbon isotope excursion.

^a*Shackleton et al.* [1984].

^b*Pospichal and Wise* [1990].

^c*Bralower and Mutterlose* [1995].

^d*Sigurdsson et al.* [1997].

Figure 3. Scanning electron microscope (SEM) images of planktonic foraminiferal inner wall structure. The right side of each image is a magnification of the left side, and the scale bars depicted in Figure 3a apply to each subsequent specimen. Figures 3a and 3b (Site 527) and 3c and 3d (Site 690) contain open pores and primary features in the wall structure and lack recrystallized calcite. By contrast, 3e and 3f (Site 1001) contain numerous rhombs of recrystallized calcite, abundant diagenetic overgrowth on the inner walls, and infilled pores.

3.2. Calcium Carbonate Records

Sites 527, 690, 999, and 1001 (Figure 6) all exhibit significant decreases in CaCO_3 in the LPTM interval [Bralower *et al.*, 1997; this study]. At Site 527, CaCO_3 contents decrease from 80% within the underlying chalk (201.14 mbsf) to <1% in the overlying homogeneous clay (201.06 mbsf); low CaCO_3 values persist for ~20 cm and then gradually increase to 78% at 200.59 mbsf (Table 2).

The Site 690 CaCO_3 decrease is more gradual and less dramatic than at the other sites, with the onset of the decrease

occurring below the CIE (Table 5). CaCO_3 contents decrease from 84 (170.81 mbsf) to 60% (170.61 mbsf) and then increase to 82% at the top of the LPTM (169.80 mbsf).

3.3. Stable Isotope Records

The Site 527 $\delta^{13}\text{C}$ records (Figure 7) show features similar to the LPTM CIE at other sites. The onset of the excursion is rapid, followed by a gradual recovery to post-LPTM $\delta^{13}\text{C}$ values that are ~1‰ lower than those below the LPTM. Within each stratigraphic horizon the three planktonic foraminiferal

Figure 4. SEM images of Site 527 claystone interval planktonic foraminifera. The right side is a magnification of the left side. The specimen demonstrates good wall structure preservation and the presence of a few zeolites.

size fractions demonstrate the predicted $\delta^{13}\text{C}$ increase with planktonic foraminiferal size [e.g., Shackleton *et al.*, 1985]. The 90-150 μm size fraction records a 2.42‰ decrease in $\delta^{13}\text{C}$ values, with the most negative excursion values occurring above the 20 cm gap in the record (Table 2). The corresponding $\delta^{13}\text{C}$ decrease recorded by the 150-250 μm size fraction is 2.59‰, with the most negative 150-250 μm size fraction carbon isotope value occurring in the only sample from the dissolution interval that contained sufficient foraminifera to analyze (Table 2).

Oxygen isotope records from Site 527 also contain an excursion in the LPTM interval (Figure 7). The 90-150 μm size fraction records a $\sim 1.25\text{‰}$ decrease in $\delta^{18}\text{O}$ from pre-LPTM values (201.00 mbsf) to the lowermost dissolution interval sample (200.80 mbsf), while the 150-250 μm size fraction

Figure 5. Sr/Ca records for Sites 527, 690, 865, and 1001 plotted versus age. The line at 55.5 Ma indicates the peak of the carbon isotope excursion (CIE) for reference.

Table 4. Planktonic Foraminiferal Sr/Ca Data

Sample	Depth, mbsf	Age, Ma	Sr/Ca, mmol mol ⁻¹
527-23-2, 55-56 cm	191.55	54.611	1.10
527-23-2, 71-72 cm	191.71	54.626	1.11
527-23-2, 86-87 cm	191.86	54.641	1.09
527-24-1, 27-28 cm	199.27	55.342	0.99
527-24-1, 49-50 cm	199.49	55.362	1.01
527-24-1, 73-74 cm	199.73	55.385	0.98
527-24-1, 96-97 cm	199.96	55.407	1.04
527-24-1, 110-111 cm	200.10	55.420	1.01
527-24-1, 131-132 cm	200.31	55.440	1.00
527-24-1, 141-142 cm	200.41	55.453	1.13
527-24-2, 4-5 cm	200.54	55.470	1.02
527-24-2, 9-10 cm	200.59	55.477	1.12
527-24-2, 12-14 cm	200.62	55.480	1.13
527-24-2, 16-18 cm	200.66	55.486	1.24
527-24-2, 60-62 cm	201.10	55.512	1.01
527-24-2, 64-66 cm	201.14	55.513	1.06
527-24-2, 68-70 cm	201.18	55.514	0.99
527-24-2, 73-75 cm	201.23	55.516	1.02
527-24-2, 80-82 cm	201.30	55.519	1.23
527-24-2, 95-97 cm	201.45	55.524	1.09
527-24-2, 106-108 cm	201.56	55.528	1.05
527-24-3, 41-42 cm	202.41	55.557	1.43
527-24-3, 73-74 cm	202.73	55.569	1.42
690B-19H-2, 118-124 cm	169.58	55.452	1.36
690B-19H-3, 40-42 cm	170.31	55.497	1.27
690B-19H-3, 51-53 cm	170.42	55.553	1.42
690B-19H-3, 60-62 cm	170.51	55.557	1.14
690B-19H-3, 66-68 cm	170.57	55.561	1.31
690B-19H-3, 74-76 cm	170.65	55.564	1.46
690B-19H-3, 115-121 cm	171.05	55.582	1.35
865C-12H-3, 120-122 cm	102.50	55.470	1.27
865C-12H-3, 140-142 cm	102.70	55.485	1.20
865C-12H-4, 4-5 cm	102.85	55.496	0.98
865C-12H-4, 6-8 cm	102.86	55.497	1.11
865C-12H-4, 10-12 cm	102.90	55.500	0.99
865C-12H-4, 16-18 cm	102.94	55.502	1.47
865C-12H-4, 20-22 cm	103.00	55.506	1.19
865C-12H-4, 30-32 cm	130.10	55.512	1.20
865C-12H-4, 40-42 cm	103.20	55.518	1.02
865C-12H-4, 70-72 cm	103.50	55.536	1.08
865C-12H-4, 90-92 cm	103.70	55.548	1.11
1001A-27R-1, 56-58 cm	237.76	55.440	0.71
1001A-27R-1, 79-81 cm	237.99	55.449	0.77
1001A-27R-1, 89-91 cm	238.09	55.453	0.65
1001A-27R-1, 96-98 cm	238.16	55.456	0.63
1001A-27R-1, 106-108 cm	238.26	55.460	0.55
1001A-27R-2, 6-7 cm	238.46	55.467	0.59
1001A-27R-2, 11-12 cm	238.51	55.469	0.58
1001A-27R-2, 15-16 cm	238.55	55.471	0.55
1001A-27R-2, 78-79 cm	239.18	55.504	0.75
1001A-27R-2, 84-85 cm	239.24	55.506	0.79
1001A-27R-3, 11-13 cm	239.66	55.516	0.77
1001A-27R-3, 22-24 cm	239.77	55.519	0.76
1001A-27R-3, 41-43 cm	239.96	55.523	0.69
1001A-27R-3, 60-62 cm	240.15	55.528	0.72

$\delta^{18}\text{O}$ values decrease by 0.94‰. Above the $\delta^{18}\text{O}$ minima (in each record), values gradually increase, and post-LPTM $\delta^{18}\text{O}$ values are higher than pre-LPTM values (Figure 7 and Table 2).

4. Discussion

4.1. Site 527 Stable Isotope Records: Preservation

Planktonic foraminiferal stable isotopes can provide information about the prevailing paleoenvironmental condi-

Figure 6. CaCO_3 records for Sites 690, 527, 999, and 1001 plotted versus age. The left to right progression represents the geographic progression from south to north. The horizontal line at 55.5 Ma indicates the position of the peak CIE for reference. The CIE within Site 527 may have occurred lower toward the base of the dissolution interval, but no foraminifera were available for stable isotopic analyses within those horizons. Sites 999 and 1001 CaCO_3 data are from Bralower *et al.* [1997], and some of the Site 690 data are from O'Connell [1990].

tions at the time of test calcification [e.g., Emiliani, 1955a]. However, recrystallization and, to a lesser extent, dissolution of calcite during burial can alter stable isotope values [e.g., Anderson and Arthur, 1983]. The upper Paleocene samples investigated were recovered from subbottom depths of several hundred meters and consequently are susceptible to diagenetic alteration. In addition, planktonic microfossils in subtropical and tropical locations such as Sites 527, 865, and 1001 are potentially more sensitive to early diagenesis because of the temperature difference between warm surface and cool deep waters [e.g., Schrag *et al.*, 1995].

Table 5. Site 690 CaCO_3 Data Generated in This Investigation

Sample	Depth, mbsf	Age, Ma	CaCO_3 , %
19H-2, 20-21 cm	168.60	55.422	95.38
19H-2, 40-41 cm	168.80	55.426	88.89
19H-2, 60-61 cm	169.00	55.431	89.87
19H-2, 80-81 cm	169.20	55.435	87.08
19H-2, 100-101 cm	169.40	55.440	78.52
19H-2, 119-120 cm	169.59	55.452	73.60
19H-2, 130-131 cm	169.70	55.459	76.04
19H-2, 140-141 cm	169.80	55.465	82.45
19H-2, 149-150 cm	169.89	55.471	76.36
19H-3, 5-6 cm	169.96	55.475	73.32
19H-3, 15-16 cm	170.06	55.482	71.41
19H-3, 25-26 cm	170.16	55.488	74.59
19H-3, 35-36 cm	170.26	55.494	68.06
19H-3, 40-41 cm	170.31	55.497	66.53
19H-3, 45-46 cm	170.36	55.550	67.14
19H-3, 50-51 cm	170.41	55.553	69.35
19H-3, 55-56 cm	170.46	55.555	68.09
19H-3, 60-61 cm	170.51	55.557	65.30
19H-3, 67-68 cm	170.58	55.561	64.05
19H-3, 70 cm	170.61	55.562	59.56
19H-3, 75-76 cm	170.66	55.564	72.97
19H-3, 80-81 cm	170.71	55.567	77.07
19H-3, 90-91 cm	170.81	55.571	83.94
19H-3, 100-101 cm	170.91	55.576	81.68
19H-3, 120-121 cm	171.11	55.585	89.35
19H-3, 140-141 cm	171.31	55.594	73.35
19H-4, 10-11 cm	171.53	55.605	88.43
19H-4, 30-31 cm	171.73	55.614	86.86

Figure 7. Site 527 planktonic foraminiferal stable isotope records plotted against stratigraphic depth for the three size fractions of *Acarinina soldadoensis* (90-150, 150-250, and 300-355 μm). The shaded band indicates the position of the claystone interval within Hole 527, Core 24, Section 2.

Sr/Ca values can be used to assess foraminiferal preservation because Sr/Ca ratios of low-Mg marine biogenic calcite are relatively high and uniform while the ratios of slowly precipitated diagenetic calcite are low and typically variable [Lewin and Chow, 1961; Lorens and Bender, 1980; Baker *et al.*, 1982]. Mean Sr/Ca values (Figure 5) for the samples used in this study indicate a higher degree of diagenetic alteration at Site 1001 relative to the other three sites. The Sr/Ca values at Sites 527, 690, and 865 are high and show <0.5 mmol mol^{-1} of intrasite variability, indicating that the Site 527 foraminifera, similar to those at Sites 690 and 865, have not been recrystallized.

SEM observations are consistent with the trends in Sr/Ca data. Foraminifera at Sites 527, 690, and 865 are well preserved (Figures 3 and 4), while Site 1001 foraminifera exhibit visual evidence of recrystallized, diagenetic calcite (Figure 3). Thus preservational data indicate that Site 527 planktonic foraminiferal stable isotope values, as well as those from Sites 690 and 865 [Kennett and Stott, 1991; Bralower *et al.*, 1995; this study], were not altered during burial. Despite the presence of zeolites inside some chambers of Site 527 LPTM foraminifera (Figure 4) the $\delta^{18}\text{O}$ values derive entirely from foraminiferal calcite because phosphoric acid digestion of the calcite for mass spectrometry is not intensive enough to liberate any zeolite oxygen [e.g., Borthwick and Harmon, 1982].

4.2 Gauging the Completeness of the Site 527 LPTM Record

The intervals of decreased CaCO_3 within the LPTM sections at Sites 527, 690, 999, and 1001 (Figure 6) indicate that corrosive intermediate and deep waters bathed the Atlantic and Caribbean basins during the LPTM. However, the presence of CaCO_3 below and above the dissolution intervals indicates that these sections were above the lysocline before and after the LPTM. The inferred transient change in deep water circulation and the oxidation of hydrate methane [e.g., Dickens *et al.*, 1995] could have increased CaCO_3 dissolution.

The completeness of deep-sea LPTM records has been questioned because of the pervasive calcium carbonate dissolution

[e.g., *Dickens*, 1998]. For example, dissolution at Site 527 has left a 20 cm section at the base of the LPTM interval almost completely barren of planktonic foraminifera. The -2.59‰ carbon isotope excursion at Site 527 is significantly less than the ~ -3 to -4‰ average excursion magnitude recorded in “expanded” LPTM sections (e.g., ODP Site 690 [Kennett and Stott, 1991], Bighorn Basin, Wyoming [Koch et al., 1992], Caravaca, Spain [Canudo et al., 1995]. Because of the absence of calcium carbonate in the basal 20 cm of the LPTM interval at Site 527, it is likely that the onset and peak of the CIE has been removed by dissolution.

4.3. Paleooceanographic Significance of the Oxygen Isotope Records

Planktonic foraminiferal $\delta^{18}\text{O}_{pf}$ values vary as a function of SST as well as the seawater oxygen isotope composition ($\delta^{18}\text{O}_{sw}$). Seawater $\delta^{18}\text{O}_{sw}$ values vary in response to a combination of processes. Globally, the $\delta^{18}\text{O}_{sw}$ is determined to a large extent by changes in ice volume and locally by variations in precipitation and evaporation (prec/evap) [see *Zachos et al.*, 1994]. Of particular interest to Paleogene paleoceanographic reconstructions is the effect of changes in prec/evap. Such changes result in an oxygen isotope “salinity effect” that is positively correlated with $\delta^{18}\text{O}_{sw}$ and consequently $\delta^{18}\text{O}_{pf}$.

Although we cannot unequivocally determine the magnitude of LPTM sea surface warming at Site 527 (Figure 7), we can constrain the range of warming by estimating the extent to which local changes in prec/evap contributed to the Site 527 $\delta^{18}\text{O}_{pf}$ shift. For global $\delta^{18}\text{O}_{sw}$, assuming there was no permanent glacial ice during the late Paleocene, we use a value of $\sim -1.0\text{‰}$ [Matthews and Poore, 1980; Miller et al., 1987]. As a late Paleocene baseline for the local $\delta^{18}\text{O}_{sw}$, we simply use the Holocene $\sim +0.8\text{‰}$ offset from mean $\delta^{18}\text{O}_{sw}$ for the eastern subtropical South Atlantic [Zachos et al., 1994], which yields a $\delta^{18}\text{O}_{sw}$ of -0.2‰ for Site 527.

From these initial parameters we independently estimate the temperature and prec/evap (salinity) changes required to produce a $\sim 0.94\text{‰}$ $\delta^{18}\text{O}_{pf}$ decrease (within the 150–250 μm size fraction record). For this exercise we use the equation of Erez and Luz [1983]:

$$T(^{\circ}\text{C}) = 16.998 + -4.52 (\delta^{18}\text{O}_{pf} - \delta^{18}\text{O}_{sw}) + 0.028 (\delta^{18}\text{O}_{pf} - \delta^{18}\text{O}_{sw})^2$$

to determine temperature (i.e., constant salinity). For the pre-LPTM $\delta^{18}\text{O}_{pf}$ value of -0.60‰ , the LPTM $\delta^{18}\text{O}_{pf}$ value of -1.54‰ , and a $\delta^{18}\text{O}_{sw}$ of -0.2‰ (at ~ 35 ppt salinity), we estimate temperatures of $\sim 16.3^{\circ}$ and $\sim 20.6^{\circ}\text{C}$, respectively, yielding a temperature increase of 4.3°C (Figure 8a).

At the other extreme, assuming constant temperature, we then calculate the salinity decrease required to generate the same change in $\delta^{18}\text{O}_{pf}$. Figure 8a depicts seawater temperature-salinity relationships with contours of $\delta^{18}\text{O}$ [after *Railsback et al.*, 1989]. The $\delta^{18}\text{O}_{pf}$ values prior to and during the peak of the LPTM are plotted on the diagram with the corresponding temperature and salinity changes required to produce the LPTM $\delta^{18}\text{O}$ decrease. Salinity would have to decrease from ~ 35.0 to ~ 32.0 ppt (while holding the SST at 16.3°C) to generate a 0.94‰ $\delta^{18}\text{O}_{pf}$ decrease.

Figure 8. Temperature (T) - salinity (S) diagrams with contours of $\delta^{18}\text{O}$ [after *Railsback et al.*, 1989; *O'Connell et al.*, 1996] for (a) Site 527 and (b) Site 690. Pre- and peak-LPTM $\delta^{18}\text{O}$ values are indicated in T - S space for both sites, with different “paths” for generating the LPTM $\delta^{18}\text{O}$ change depicted. T indicates that temperature was the only effect on $\delta^{18}\text{O}$ values, S reflects solely a salinity change, and T-S indicates that both effects were considered. Site 690 $\delta^{18}\text{O}$ data and paleotemperature reconstructions are from *Kennett and Stott* [1991] and *Stott et al.* [1990]. The depicted $\sim 9^{\circ}\text{C}$ T change at Site 690 reflects the use of a different paleotemperature equation in this investigation.

A 3 ppt decrease in sea surface salinity in an open ocean location is unrealistic given that the total modern range of South Atlantic open ocean salinities is only ~ 2 - 3 ppt [e.g., *Defant*, 1961]. Yet, it is also unrealistic to assume that a $\sim 4.3^{\circ}\text{C}$ temperature change was not accompanied by a change in prec/evap. In fact, model simulations of global climate predict a more active hydrologic cycle and altered prec/evap patterns during the LPTM [O'Connell et al., 1996; Bice et al., 1997; Sloan and Thomas, 1999]. Regionally, an increase in prec/evap may have contributed to the decrease in $\delta^{18}\text{O}_{pf}$ values.

If salinity did decrease within the subtropical South Atlantic during the LPTM, it was probably no more than 2 ppt, which is the entire open ocean range for the modern South Atlantic. This assertion is supported to some extent by the work of *Barron and Peterson* [1990], who found regional-scale variations in open ocean salinities of ~ 1 - 1.5 ppt in model simulations of mid-Cretaceous climate/ocean circulation. If we accept 2 ppt as the maximum possible salinity increase, the minimum amount of sea surface warming can be constrained. As shown in Figure 8a, the maximum possible 2

ppt salinity decrease corresponds to a $\sim 0.8\text{‰}$ $\delta^{18}\text{O}_{\text{sw}}$ change [Zachos et al., 1994]. The LPTM $\delta^{18}\text{O}_{\text{pf}}$ temperature effect resulting from a 2 ppt decrease in sea surface salinity is $\sim -0.16\text{‰}$ (-0.94‰ to -0.8‰), corresponding to a 0.7°C warming (from $\sim 16.3^{\circ}$ to 17.0°C). The range of estimated subtropical South Atlantic sea surface warming is therefore 0.7° - 4.3°C . Because of the dissolution gap in the record, this range should be viewed as a minimum estimate of the extent of subtropical warming.

Though somewhat speculative, it is useful to consider the role of salinity change on estimates of LPTM warming elsewhere. For example, we can employ a similar approach to determine the possible effect of a prec/evap change on the LPTM planktonic oxygen isotope record at Southern Ocean ODP Site 690. Robert and Kennett [1994] found that the kaolinite content at Site 690 increased across the LPTM and attributed the change in clay mineralogy to an altered weathering regime. While the kaolinite increase could have resulted from numerous other factors (e.g., altered erosional patterns or sedimentation rates), the possibility that the kaolinite increase reflects a significant prec/evap change within the Antarctic and Southern Ocean region warrants consideration of a salinity effect on the Site 690 oxygen isotope record. The $\delta^{18}\text{O}_{\text{pf}}$ decrease of $\sim 2.0\text{‰}$ (from -0.2 to -2.2‰) was attributed entirely to $\sim 7^{\circ}$ - 8°C of surface water warming [Kennett and Stott, 1991] (Figure 8b). If we assume a 2 ppt salinity decrease and the consequent $\delta^{18}\text{O}$ salinity effect of 0.8‰ , yielding a 1.2‰ $\delta^{18}\text{O}$ temperature effect, we estimate a SST increase of 5.3°C at Site 690 (Figure 8b). The original 7° - 8°C estimate [Kennett and Stott, 1991] assumed no change in salinity. The prec/evap effect at Site 690 may have been even larger than estimated above possibly because of significantly lower high-latitude meteoric water $\delta^{18}\text{O}$ values [e.g., Bice et al., 1996].

4.4. Implications of Subtropical Warming During the LPTM

The isotopic data from Site 527 represent the first evidence for subtropical SST warming during the LPTM, a finding that has important implications for the proposed cause(s) of the event. Model simulations of increased atmospheric CO_2 con-

centrations predict warming at all latitudes [Sloan and Thomas, 1999], thus the new evidence for subtropical sea surface warming strengthens the likelihood that LPTM warming may have resulted from enhanced atmospheric CO_2 concentrations. Possible sources of an atmospheric CO_2 increase during the late Paleocene include intense NAIP volcanism [e.g., Eldholm and Thomas, 1993] and increased seafloor hydrothermal activity [e.g., Rea et al., 1990]. Potentially superimposed on these processes is the possibility of rapid, short-term CO_2 input from the oxidation of methane from dissociated hydrates [Dickens et al., 1995, 1997].

In light of the new evidence for subtropical Atlantic sea surface warming the lack of stable isotopic evidence for warming at subtropical and tropical Pacific Sites 47.2 and 865 [Stott, 1992; Bralower et al., 1995] may have been the result of incomplete sections (similar to Site 527) in which the peak of the LPTM was not recorded by the sedimentary sequence or, alternatively, may have resulted from a salinity increase that could have muted the $\delta^{18}\text{O}_{\text{pf}}$ temperature response. In addition, the decrease in Caribbean Site 1001 bulk carbonate $\delta^{18}\text{O}$ values may indicate tropical warming [Bralower et al., 1997], although foraminiferal diagenesis precludes paleotemperature estimation.

5. Conclusions

Preservation data (SEM and Sr/Ca) indicate that planktonic foraminifera from Sites 527, 690, and 865 are not significantly altered by diagenesis. The Site 527 $\delta^{18}\text{O}$ record is evidence for warming of the eastern subtropical South Atlantic during the LPTM; the constraints on salinity variations suggest warming ranged from 0.7° to 4.3°C . The evidence for subtropical warming increases the likelihood that enhanced atmospheric CO_2 concentration ultimately was responsible for the LPTM warming.

Acknowledgments. We would like to extend our gratitude to Carol Arnosti, Gary Dwyer, Allen Glazner, Charlie Jones, Clay Kelly, Conrad Neumann, Charlie Paull, and Bill Ussler for technical assistance and insightful discussions. In addition, we thank Ken Miller, Ben Cramer, and Ellen Thomas for thoughtful and constructive reviews. This work was supported by a GSA Student Grant to DJT and a JOI/USSAC grant to TJB.

References

- Anderson, T.F. and M.A. Arthur, Stable isotopes of oxygen and carbon and the application to sedimentologic and paleoenvironmental problems, in *Stable Isotopes in Sedimentary Geology*, edited by M.A. Arthur et al., *SEPM Short Course*, 10, 1-151, 1983.
- Aubry, M.-P., From chronology to stratigraphy: Assessing the continuity of the stratigraphic record, in *Geochronology, Time Scales and Global Stratigraphic Correlations: Framework for an Historical Geology*, edited by W.A. Berggren, et al., *SEPM Spec. Publ.* 54, 213-275, 1995.
- Aubry, M.-P., W.A. Berggren, D.V. Kent, J.J. Flynn, K.D. Klitgord, J.D. Obradovich, and D.R. Prothero, Paleogene geochronology: An integrated approach, *Paleoceanography*, 3, 707-742, 1988.
- Baker, P.A., J.M. Gieskes, and H. Elderfield, Diagenesis of carbonates in deep-sea sediments: evidence from Sr/Ca ratios and interstitial dissolved Sr^{+2} data, *Sediment. Petrol.*, 52, 71-82, 1982.
- Barron, E.J., and W.H. Peterson, Mid-Cretaceous ocean circulation: Results from model sensitivity studies, *Paleoceanography*, 5, 319-337, 1990.
- Berggren, W.A., D.V. Kent, C.C. Swisher III, and M.-P. Aubry, A revised Cenozoic geochronology and chronostratigraphy, in *Geochronology, Time Scales and Global Stratigraphic Correlations: Framework for an Historical Geology*, edited by W.A. Berggren et al., *SEPM Spec. Publ.* 54, 129-212, 1995.
- Bice, K.L., M.A. Arthur, and L. Marinovich Jr., Late Paleocene Arctic Ocean shallow-marine temperatures from mollusc stable isotopes, *Paleoceanography*, 11, 241-249, 1996.
- Bice, K.L., E.J. Barron, and W.H. Peterson, Continental runoff and early Cenozoic bottom-water sources, *Geology*, 25, 951-954, 1997.
- Borthwick, J., and R.S. Harmon, A note regarding ClF_3 as an alternative to BrF_3 for oxygen isotope analysis, *Geochim. Cosmochim. Acta*, 27, 1665-1668, 1963.
- Boyle, E.A., Cadmium, zinc, copper, and barium in foraminifera tests, *Earth Planet. Sci. Lett.*, 53, 11-35, 1981.
- Bralower, T.J., and J. Mutterlose, Calcareous nannofossil biostratigraphy of ODP Site 865, Allison Guyot, Central Pacific Ocean: a tropical Paleogene reference section, *Proc. Ocean Drill. Program Sci. Res.*, 143, 31-74, 1995.
- Bralower, T.J., J.C. Zachos, E. Thomas, M. Parrow, C.K. Paull, D.C. Kelly, I. Premoli Silva, W.V. Sliter, and K.C. Lohmann, Late Paleocene to Eocene paleoceanography of the equatorial Pacific Ocean: Stable isotopes recorded at Ocean Drilling Program Site 865, Allison Guyot, *Paleoceanography*, 10, 841-865, 1995.
- Bralower, T.J., D.J. Thomas, J.C. Zachos, M.M. Hirschmann, U. Rohl, H. Sigurdsson, E. Thomas, and D.L. Whitney, High-resolution records of late Paleocene thermal maximum and circum-Caribbean volcanism: Is there a causal link?, *Geology*, 25, 963-966, 1997.
- Canudo, J.I., G. Keller, E. Molina, and N. Ortiz, Planktonic foraminiferal turnover and $\delta^{13}\text{C}$ isotopes across the Paleocene-Eocene transition at Caravaca and Zumaya, Spain, *Paleogeogr. Palaeoclimatol. Palaeoecol.*, 114, 75-100, 1995.
- Charisi, S.D. and B. Schmitz, Stable ($\text{d}13\text{C}$, $\text{d}18\text{O}$)

- and strontium ($^{87}\text{Sr}/^{86}\text{Sr}$) isotopes through the Paleocene at Gehel Aweina, eastern Tethyan region, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 116, 103-129, 1995.
- Defant, S., *Physical Oceanography* vol. 1, Pergamon, Tarrytown, N.Y., 1961.
- Dickens, G.R., Comment, *Geology*, 26, 670, 1998.
- Dickens, G.R., J.R. O'Neil, D.C. Rea, and R.M. Owen, Dissociation of oceanic methane hydrate as a cause of the carbon isotope excursion at the end of the Paleocene, *Paleoceanography*, 10, 965-971, 1995.
- Dickens, G.R., M.M. Castillo, and J.C.G. Walker, A blast of gas in the latest Paleocene: simulating first-order effects of massive dissociation of oceanic methane hydrate, *Geology*, 25, 258-262, 1997.
- Eldholm, O., and E. Thomas, Environmental impact of volcanic margin formation, *Earth Planet. Sci. Lett.*, 117, 319-329, 1993.
- Emiliani, C., Pleistocene temperatures, *J. Geology*, 63, 538-578, 1955.
- Erez, J., and B. Luz, Experimental paleotemperature equation for planktonic foraminifera, *Geochim. Cosmochim. Acta*, 47, 1025-1031, 1983.
- Hooker, J.J., Mammalian biostratigraphy across the Paleocene-Eocene boundary in the Paris, London and Belgian basins, in *Correlation of the Early Paleogene in Northwest Europe*, *Geol. Soc. Spec. Publ.* 101, edited by R.W.O.B. Knox, R.M. Corfield, and R.E. Dunay, 205-218, 1996.
- Kelly, D.C., T.J. Bralower, J.C. Zachos, I. Premoli Silva, and E. Thomas, Rapid diversification of planktonic foraminifera in the tropical Pacific (ODP Site 865) during the late Paleocene thermal maximum, *Geology*, 24, 423-426, 1996.
- Kelly, D.C., T.J. Bralower, and J.C. Zachos, Evolutionary consequences of the latest Paleocene thermal maximum for tropical planktonic foraminifera, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 141, 139-161, 1998.
- Kennett, J. P., and L.D. Stott, Proteus and Proto-Oceanus: ancestral Paleogene oceans as revealed from Antarctic stable isotopic results: ODP Leg 113, *Proc. Ocean Drill. Program Sci. Results*, 113, 865-880, 1990.
- Kennett, J. P., and L.D. Stott, Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Palaeocene, *Nature*, 353, 225-229, 1991.
- Koch, P.L., J.C. Zachos, and P.D. Gingerich, Correlation between isotope records in marine and continental carbon reservoirs near the Palaeocene/Eocene boundary, *Nature*, 358, 319-322, 1992.
- Lewin, R.A., and T.J. Chow, La enpreno de strontio en Kokolitoroj, *Plant Cell Physiol.*, 2, 203-208, 1961.
- Lorens, R.B., and M.L. Bender, The impact of solution chemistry on *Mytilus edulis* calcite and aragonite, *Geochim. Cosmochim. Acta*, 44, 1265-1278, 1980.
- Matthews, R.K., and R.Z. Poore, Tertiary $\delta^{18}\text{O}$ record and glacio-eustatic sea-level fluctuations, *Geology*, 8, 501-504, 1980.
- Miller, K.G., R.G. Fairbanks, and G.S. Mountain, Tertiary oxygen isotope synthesis, sea-level history/continental margin erosion, *Paleoceanography*, 2, 1-19, 1987b.
- O'Connell, S.B., Variations in upper Cretaceous and Cenozoic calcium carbonate percentages, Maud Rise, Weddell Sea, Antarctica, *Proc. Ocean Drill. Program Sci. Results*, 113, 971-984, 1990.
- O'Connell, S.B., M.A. Chandler, and R. Rucy, Implications for the creation of warm saline deep water: Late Paleocene reconstructions and global climate model simulations, *Geol. Soc. Amer. Bull.*, 108, 270-284, 1996.
- Pospichal, J.J., and S.W. Wise Jr., Paleocene to Middle Eocene calcareous nannofossils of ODP Sites 689 and 690, Maud Rise, Weddell Sea, *Proc. Ocean Drill. Program Sci. Results*, 113, 613-631, 1990.
- Railsback, L.B., T.F. Anderson, S.C. Ackerly, and J.L. Cisne, Paleocceanographic modeling of temperature-salinity profiles from stable isotopic data, *Paleoceanography*, 4, 585-591, 1989.
- Rea, D.K., J.C. Zachos, R.M. Owen, and P.D. Gingerich, Global change at the Paleocene-Eocene boundary: Climatic and evolutionary consequences of tectonic events, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 79, 117-128, 1990.
- Robert, C., and J.P. Kennett, Antarctic subtropical humid episode at the Paleocene-Eocene boundary: Clay mineral evidence, *Geology*, 22, 211-214, 1994.
- Schmitz, B., F. Asaro, E. Molina, S. Monechi, K. von Salis, and R.P. Speijer, High-resolution iridium, $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, foraminifera and nannofossil profiles across the latest Paleocene benthic extinction event at Zumaya, Spain, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 133, 49-68, 1997.
- Schrag, D.P., D.J. DePaolo, and F.M. Richter, Reconstructing past sea surface temperatures: Correcting for diagenesis of bulk marine carbonate, *Geochim. Cosmochim. Acta*, 59, 2265-2278, 1995.
- Shackleton, N.J., M.A. Hall, and A. Boersma, Oxygen and carbon isotope data from Leg 74 foraminifera, *Initial Rep. Deep Sea Drill. Project*, 74, 599-612, 1984.
- Shackleton, N.J., R.M. Corfield, and M.A. Hall, Stable isotope data and the ontogeny of Paleocene planktonic foraminifera, *J. Foraminiferal Res.*, 15, 321-336, 1985.
- Sigurdsson, H., R.M. Leckie, and G. Acton, *Proceedings of the Ocean Drilling Program: Initial Reports*, vol. 165, Ocean Drill. Program, College Station, Tex., 1997.
- Sloan, L.C., and E. Thomas, Global climate of the late Paleocene: Modeling the circumstances associated with a climatic "event," in *The Paleocene/Eocene boundary*, edited by M.-P. Aubry et al., Lamont-Doherty Earth Observatory, Palisades, N.Y., 1999.
- Stott, L.D., Higher temperatures and lower oceanic $p\text{CO}_2$: A climatic enigma at the end of the Paleocene epoch, *Paleoceanography*, 7, 395-404, 1992.
- Stott, L.D., J.P. Kennett, N.J. Shackleton, and R.M. Corfield, The evolution of Antarctic surface waters during the Paleogene: Inferences from the stable isotopic compositions of planktonic foraminifera, ODP Leg 113, *Proc. Ocean Drill. Program Sci. Results*, 113, 849-863, 1990.
- Thomas, E., Late Cretaceous-early Eocene mass extinctions in the deep-sea, in *Global Catastrophes*, edited by V.L. Sharpton, and P.D. Ward, *Spec. Pap. Geol. Soc. Am.*, 247, 481-496, 1990.
- Thomas, E., and N.J. Shackleton, The Paleocene-Eocene benthic foraminiferal extinction and stable isotope anomalies, in *Correlation of the Early Paleogene in Northwest Europe*, *Geol. Soc. Spec. Publ.* 101, edited by R.W.O.B. Knox, R.M. Corfield, and R.E. Dunay, 401-441, 1996.
- Tjalsma, R.C., and G.P. Lohmann, Paleocene-Eocene bathyal and abyssal benthic foraminifera from the Atlantic Ocean, *Micropaleontol. Spec. Publ.* 4, 90, 1983.
- Wing, S.L., T.M. Bown, and J.D. Obradovich, Early Eocene biotic and climatic change in interior western America, *Geology*, 19, 1189-1192, 1991.
- Zachos, J.C., K.C. Lohmann, J.C.G. Walker, and S.W. Wise, Abrupt climate change and transient climates during the Paleogene: A marine perspective, *J. Geol.*, 101, 191-213, 1993.
- Zachos, J.C., L.D. Stott, and K.C. Lohmann, Evolution of early Cenozoic marine temperatures, *Paleoceanography*, 9, 353-387, 1994.

T.J. Bralower and D. J. Thomas, Department of Geological Sciences, University of North Carolina, Chapel Hill, NC 27599-3315. (dthomas1@email.unc.edu)

J.C. Zachos, Department of Earth Sciences, University of California, Santa Cruz, CA 95064.

(Received December 10, 1998;
revised May 20, 1999;
accepted May 25, 1999.)