High-resolution records of the late Paleocene thermal maximum and circum-Caribbean volcanism: Is there a causal link?

T. J. Bralower
D. J. Thomas — Department of Geology, University of North Carolina, Chapel Hill, North Carolina 27599
J. C. Zachos Earth Science Department, University of California, Santa Cruz, California 95064
M. M. Hirschmann* Department of Geology, University of North Carolina, Chapel Hill, North Carolina 27599
U. Röhl Geosciences Department, Bremen University, D-28334 Bremen, Germany
H. Sigurdsson Graduate School of Oceanography, University of Rhode Island, Narragansett, Rhode Island 02882
E. Thomas Department of Earth and Environmental Sciences, Wesleyan University, Middletown, Connecticut 06457
D. L. Whitney* Department of Geology, University of North Carolina, Chapel Hill, North Carolina 27599

ABSTRACT

Two recently drilled Caribbean sites contain expanded sedimentary records of the late Paleocene thermal maximum, a dramatic global warming event that occurred at ca. 55 Ma. The records document significant environmental changes, including deep-water oxygen deficiency and a mass extinction of deep-sea fauna, intertwined with evidence for a major episode of explosive volcanism. We postulate that this volcanism initiated a reordering of ocean circulation that resulted in rapid global warming and dramatic changes in the Earth's environment.

INTRODUCTION

The abrupt warming that took place in the late Paleocene Epoch (55 Ma) is one of the most pronounced, transient ($<10^5$ yr) climatic events in the geologic record (e.g., Zachos et al., 1993). Known as the late Paleocene thermal maximum (LPTM), this event was associated with dramatic changes in the Earth's oceans, climate, and biosphere. High-latitude sea-surface temperatures (SSTs) rose by ~6 °C, and the intensity of atmospheric circulation diminished (e.g., Rea et al., 1990). The contribution from low-latitude sources to deep waters (we use the term "deep waters" to refer to all waters beneath the thermocline) increased, with a concomitant warming of ~8 °C (e.g., Kennett and Stott, 1991). Warming of deep waters may have contributed to the most severe mass extinction of deep-sea benthic foraminifers in the past 100 m.y. (see references in Thomas and Shackleton, 1996). Rapid rates of speciation are observed in fossil groups such as the planktic foraminifers and land mammals (see references in Rea et al., 1990). The LPTM correlates with a large (~3‰), and abrupt, negative excursion in the stable carbon isotope composition of marine and terrestrial materials (e.g., Kennett and Stott, 1991; Koch et al., 1992). The only feasible proposed mechanism capable of causing such a large and rapid change in carbon cycling is dissociation of methane hydrates as a consequence of the warming of deep waters (Dickens et al., 1995).

The effects of the LPTM are now well documented, but its ultimate cause has remained elusive. It has been postulated that volcanism in the North Atlantic igneous province warmed high-latitude climate and switched the dominant source of deep waters to low latitudes (e.g., Rea et al., 1990; Eldholm and Thomas, 1993). However, climate models (Sloan and Thomas, 1997) combined with the lack of observed tropical warming in the LPTM (e.g., Stott, 1992) are at odds with a volcanic-CO_2-induced warming mechanism.

Here we describe a remarkable environmental and volcanic record of the LPTM from the Caribbean Sea. We first discuss changes in climate and deep-water oxygenation as inferred from stable isotope and other data, and then propose a possible causal connection between explosive circum-Caribbean volcanism and the LPTM.

LATE PALEOCENE THERMAL MAXIMUM RECORDS FROM THE CARIBBEAN

Late Paleocene thermal maximum records were recovered during Ocean Drilling Program Leg 165 in the Caribbean Sea at site 999 (Colombia Basin) and site 1001 (lower Nicaraguan Rise) (Fig. 1). The sites were located at ~10°N in the late Paleocene (Sigurdsson et al., 1997). At sites 999 and 1001, the event is recorded by a claystone layer that is 0.42 and 0.80 m thick, respectively. At site 999, the claystone shows faint lamination in some intervals (Fig. 2). Both records indicate diminished bioturbation throughout the LPTM.

18 multicolored, 1.5- to 13-cm-thick volcanic tephra layers are interbedded in the upper Paleocene-lower Eocene sequence at site 1001; four (layers F, G, H, and I) lie within the claystone, and one (layer J) is directly beneath it. Three thin (1–5 mm) tephra layers (layers 1, 2, and 3; Fig. 2) occur in the claystone at site 999.

Samples were taken every 1 to 2 cm in the LPTM interval and every 10 cm within 0.5 m of the event. Calcium carbonate and organic carbon (C_{org}) contents were measured on powdered samples using a coulometer. Bulk carbonate isotope measurements on powdered samples were conducted with an autocarb device coupled to a Fisons Prism gas-source mass spectrometer at the University of California–Santa Cruz. Average precision, as determined from replicate analyses of NBS-19 and Carrera Marble carbonate standards, was better than ±0.10 for both $\delta^{18}O$ and $\delta^{13}C$ values. Minerals in the tephra layers were identified from their optical properties supplemented by scanning-electron-microscope–energy-dispersive spectrometry (EDS) and EDS and wavelength-dispersive spectrometry–electron-microprobe analyses.

INTERPRETATION OF ISOTOPE, FAUNAL, AND MINERAL DATA

Isotope analyses of bulk carbonate show an ~12‰ negative $\delta^{13}C$ excursion at site 999 and an

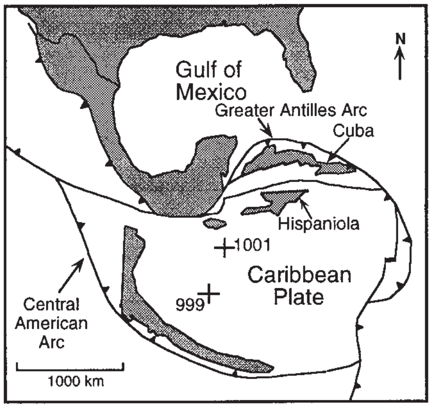

Figure 1. Tectonic reconstruction of late Paleocene Caribbean from Pindell and Barrett (1990), showing locations of sites 999 and 1001.

---

*Present address: Department of Geology and Geophysics, University of Minnesota, Minneapolis, Minnesota 55455.

Data Repository item 9760 contains additional material related to this article.

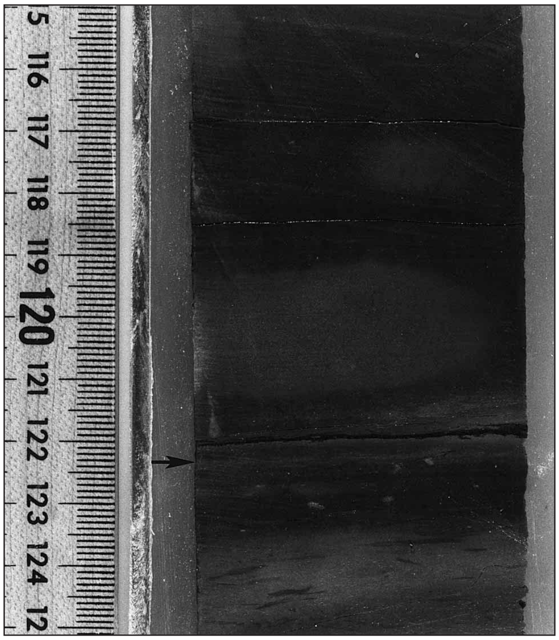

Figure 2. Photograph of lower part of LPTM at site 999. Evidence of bioturbation is seen at 123.5–124.5 cm. Tephra layer 3 (1.5 mm thick) is at 122.3 cm (arrow), and lowermost evidence for lamination is at 121.3 cm. (Note curved saw-blade marks on left above 124 cm and on right above 118 cm.)

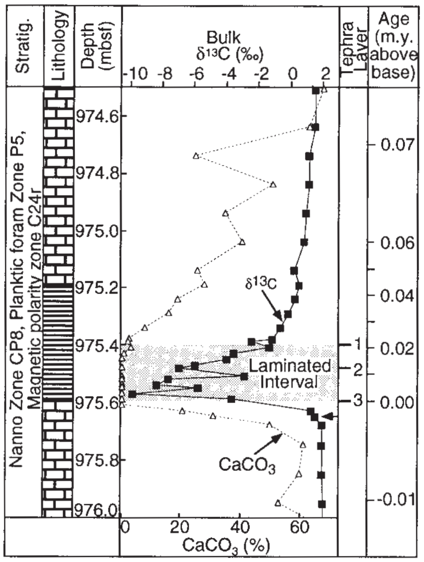

Figure 3. Carbon isotope stratigraphy of LPTM interval at site 999 (Hole 999B) plotted in meters below sea floor. Stratigraphic levels of tephra layers are shown. Level of uppermost pre-extinction benthic foraminiferal assemblages is shown by arrow at right. Biostratigraphy and magnetostratigraphic units in "Stratig." column are documented in Sigurdsson et al. (1997). Sample ages within $\delta^{13}C$ excursion interval are estimated by assuming constant sedimentation rates and age model of Thomas and Shackleton (1996) for site 690 (Maud Rise), which gives an age of 55.50 Ma for base of excursion and 55.44 Ma for horizon at which $\delta^{13}C$ values level out. Above and below excursion interval, ages were estimated assuming constant sedimentation rates for nannofossil Zone CP8 (32 m/m.y. for site 999 and 38 m/m.y. for site 1001; Sigurdsson et al., 1997). Isotope and $CaCO_3$ data are archived in the GSA Data Repository (see footnote 1 in text).

~3‰ shift at site 1001 (Figs. 3[1] and 4). The $\delta^{13}C$ excursions possess a shape similar to that of records at other sites; the onset of the excursion is sharp, minimum $\delta^{13}C$ values are reached less than 10 cm above the base of the excursion, and the recovery is gradual, occurring over 0.5 m at site 999 and 1.5 to 2.5 m at site 1001 (Figs. 3 and 4). The lengths of the $\delta^{13}C$ excursions suggest that these are among the most expanded LPTM sections available.

The LPTM interval at site 999 is ~975 m below sea floor (mbsf), and the interval at site 1001 is currently 240 mbsf, but comparison of induration with nearby site 152 suggests previous burial of ~400 m. Studies of the effects of diagenesis (e.g., Anderson and Arthur, 1983) indicate that the general direction of the $\delta^{13}C$ excursions at both sites is probably preserved since C_{org} contents are insignificant (<0.1%). The magnitude of the $\delta^{13}C$ shift at site 999, however, might be large due to the formation of ^{13}C-depleted carbonate in the claystone from early diagenesis. As with most pelagic sediments, burial diagenesis at depth has probably shifted the bulk $\delta^{18}O$ values though the relative sense of bulk $\delta^{18}O$ fluctuations, particularly at site 1001, is probably preserved (e.g., Schrag et al., 1995). The irregular, sub-horizontal base of the claystone at site 999 (Fig. 2) suggests diffusive dissolution of the upper few cm of seafloor carbonates.

There are no calcareous foraminifers in the claystones, which prevented detection of the benthic foraminiferal extinction to the centimeter level (e.g., Thomas and Shackleton, 1996). Pre-extinction benthic foraminiferal faunas were observed in samples at 975.65 mbsf in site 999 and at 3.66 m in the composite section (mcs) in site 1001 (Figs. 3 and 4). Benthic foraminiferal assemblages suggest lower-bathyal to upper-abyssal paleodepths for both sites (1500–2500 m); site 1001 was toward the lower end of that range.

Tephra layers at site 1001 consist of a smectite clay matrix and abundant phenocrysts of oscillatory-zoned sodic ($An_{\sim30}$) plagioclase, with common hornblende and biotite, as well as sanidine and ilmenite. We were unable to identify fresh glass shards, but rhyodacitic melt inclusions are preserved in some plagioclase phenocrysts. On the basis of the mineralogy of the tephra layers and the composition of the melt inclusions, we infer that these layers represent eruptions from an evolved, calc-alkalic, magmatic system. Although the thin tephras in the LPTM interval at site 999 are almost entirely altered to clay, abundant plagioclase phenocrysts were observed in layer 3. The coring gap at the base of the $\delta^{13}C$ excursion at site 1001 prohibits a definitive correlation between layer J and layer 3 at site 999. We believe that these layers are correlative because (1) they have a similar position relative to the $\delta^{13}C$ excursion; (2) the upper part of tephra layer J is bioturbated suggesting that it was deposited in predysoxic conditions; (3) sample 1001B-27R-2, 60–61 cm, in the bioturbated part of tephra layer J, has excursion $\delta^{13}C$ values; and (4) formation microscanner logs indicate that the gap between tephra J and the underlying chalk is less than 10 cm.

THE CASE FOR TROPICAL WARMING IN THE LATEST PALEOCENE

Our records yield valuable information about the effects of the LPTM on the oceanic environment. That the records are from tropical sites adds to their significance; most LPTM records are derived from temperate and high-latitude areas. At site 1001, the $\delta^{13}C$ excursion that marks the LPTM coincides with a small (0.25‰–0.5‰) but significant decrease in $\delta^{18}O$ values that begins ~0.5 m lower (Fig. 4). Because carbonate at site 1001 is dominated by planktic foraminifers and nannofossils, the decrease in bulk $\delta^{18}O$ values indicates a minor rise in tropical SSTs just prior to and during the LPTM, the first evidence for tropical warming.

LATE PALEOCENE THERMAL MAXIMUM DEEP-WATER ENVIRONMENTS

Diminished carbonate contents in sites 999 and 1001 claystones are thought to reflect shoaling of the lysocline and CCD (calcite compensation depth), similar to the records from several other LPTM sections (see references in Thomas and Shackleton, 1996). Dissolution of carbonate is consistent with addition of massive amounts of carbon from oxidized methane hydrate (e.g., Dickens et al., 1995).

---

[1] GSA Data Repository item 9760, isotope and $CaCO_3$ data, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301. E-mail: editing@geosociety.org.

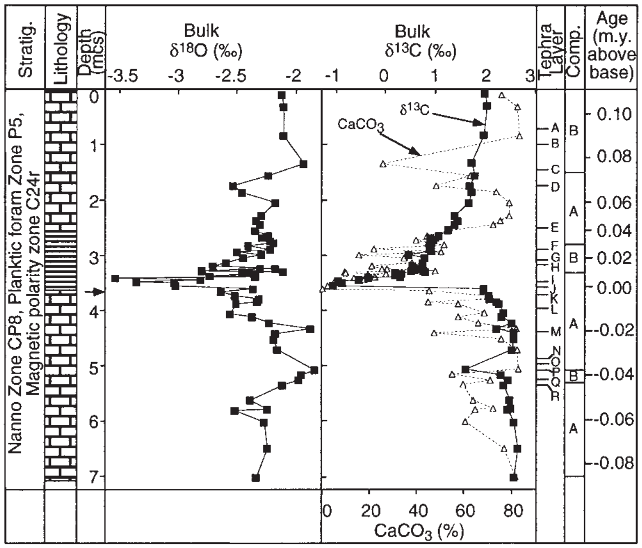

Figure 4. Carbon and oxygen isotope stratigraphy of LPTM section at site 1001. Levels of tephra layers are shown. "Comp." refers to composite section in holes 1001A and 1001B by using tephra layers (mcs refers to meters in composite section). Level of uppermost preextinction benthic foraminiferal assemblages is shown with arrow. See Figure 3 for explanation of stratigraphy.

Faint lamination in the LPTM interval at site 999 is clear evidence for dysoxic (<0.2 mL/L dissolved O_2) deep waters. Similar sedimentary structures were described in LPTM records from epicontinental seas (e.g., Kaiho et al., 1996) and site 690 (Maud Rise, South Atlantic) (Kennett and Stott, 1991), but the latter observations have been questioned by Thomas and Shackleton (1996). At site 999, the transition from oxygenated conditions, as evidenced by clearly bioturbated sedimentary rocks, to dysoxic conditions occurs over 2 cm (Fig. 2), the equivalent of 2 k.y. Well-defined lamination suggests that peak oxygen deficiency occurred some 6 k.y. after the onset of the LPTM and that conditions became steadily more oxygenated thereafter.

RECORD OF LATE PALEOCENE CIRCUM-CARIBBEAN VOLCANISM

Tephra layers at site 1001 record a volcanic episode at the Central American arc to the west or the proto-Greater Antilles arc to the north (Fig. 1), beginning in the late Paleocene (ca. 56 Ma) and continuing into the Eocene (Sigurdsson et al., 1997) (Fig. 5). The thickness of the tephra layers (up to 13 cm) combined with the distance of the sites from the proposed arcs (~1000 km) suggests that the eruptions were powerful, with volumes comparable to the 75 ka Toba eruption, the largest known late Quaternary eruption (e.g., Rampino and Self, 1992).

The records at sites 999 and 1001 allow us to draw precise time lines between circum-Caribbean volcanism and the LPTM (Fig. 5). Volcanism gradually increased in the period between 56.0 and 55.55 Ma with at least 12 eruptions recorded at site 1001. Activity intensified significantly in the 50 k.y. prior to the LPTM; at least eight tephra layers are found in this interval (Fig. 4). Tephra layer J at site 1001 and layer 3 at site 999 occur at the base of the LPTM interval as defined by the $\delta^{13}C$ excursion (i.e., just below the lowermost measured excursion $\delta^{13}C$ value). The site 999 record graphically shows that the volcanic event lies directly between sedimentary rocks that are bioturbated and those whose lamination indicates dysoxic bottom waters (Fig. 2).

CAUSAL SCENARIO FOR THE LATE PALEOCENE THERMAL MAXIMUM

The apparent synchroneity between the circum-Caribbean eruption and the onset of the LPTM suggests that there could be a causal link between volcanism and rapid climatic change. In this section, we speculate on one scenario of how a volcanic trigger might have worked.

Warming from North Atlantic Igneous Province Volcanism: Conditioning the Ocean

Large-scale North Atlantic igneous province volcanism may have set the stage for the LPTM by causing warming concentrated at high latitudes (e.g., Rea et al., 1990; Eldholm and Thomas, 1993). This volcanic activity began at ca. 61 Ma, but most of the province was formed between 54 to 57 Ma during Chrons C25r to C24r (Storey et al., 1996). A precise cooling age for the Skaergaard intrusion (55.65 ± 0.30 Ma; Hirschmann et al., 1997) combined with stratigraphic relations in E. Greenland suggests that a substantial portion of the North Atlantic igneous province flood basalts had erupted prior to the LPTM.

Effusive eruptions, such as those in the North Atlantic igneous province, have the potential to cause long-term warming because they commonly involve voluminous CO_2 degassing. Modern-day volcanic activity has not resulted in warming because the huge atmospheric-oceanic-terrestrial CO_2 reservoir negates the potential radiative greenhouse effect of degassed CO_2 (e.g., Varekamp et al., 1992). However, North Atlantic igneous province activity probably had a significant effect on late Paleocene climate because of the immense scale—yet pulsed nature—of the eruptions (e.g., Eldholm and Thomas, 1993). The presumably huge volume of CO_2 emission and the reduced CO_2 solubility in the warm, late Paleocene oceans may have enhanced accumulation in the atmosphere (e.g., Owen and Rea, 1985), creating the observed pre–LPTM warming.

Cooling from Caribbean Eruptions: Reordering Ocean Circulation

The circum-Caribbean volcanic episode was superimposed on this trend of long-term warming. The climatic effects of the circum-Caribbean eruptions were apparently too ephemeral to be preserved in marine isotope records, and thus we can only constrain them indirectly. Historically, arc volcanic eruptions have commonly caused short-term (1–10 yr) atmospheric cooling, mainly as a result of the formation of stratospheric sulfate aerosols that backscatter short-wave solar radiation (e.g., Rampino, 1991). The radiative loss resulting from the 1991 eruption of Mt. Pinatubo, at a comparable latitude to the circum-Caribbean volcanoes, was greatest in cloud-free, low-latitude oceanic areas (Minnis et al., 1993). This eruption may have been responsible for a <1 °C decrease of tropical western Pacific SSTs (Gagan and Chivas, 1995). Ice core records show that the quantity and residence time of aerosols from the Toba eruption were theoretically sufficient to cause SSTs to decrease (e.g., Zielinski et al., 1996). Thus we postulate that over short-term time scales, the circum-Caribbean eruptions slowed the long-term late Paleocene warming trend preferentially at low latitudes and therefore decreased the difference between high- and low-latitude SSTs, amplifying the climatic effect of North Atlantic igneous province volcanism.

If the difference between high- and low-latitude SSTs was reduced, low-latitude surface waters may have become denser than those at high latitudes, possibly leading to focused low-latitude downwelling and a reordering of ocean circulation (e.g., Kennett and Stott, 1991). Late Paleocene

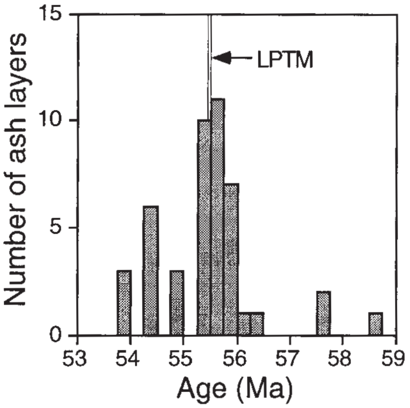

Figure 5. Frequency of tephra layers recovered at site 1001 in 0.25 m.y. increments through upper Paleocene and lower Eocene. Gaps indicate intervals of time for which no section was recovered.

general-circulation-model results show large tropical and subtropical oceanic areas with net evaporation (O'Connell et al., 1996; Sloan and Thomas, 1997). Uncoupled early Cenozoic atmospheric and oceanic general-circulation-model results show that the density contrast between competing high-latitude and subtropical deep-water sources was small (Bice, 1997). Thus even a minor additional decrease in the SST difference between subtropical and high-latitude waters could have reversed their relative buoyancies. We speculate that the eruption at the onset of the LPTM caused this buoyancy threshold to be crossed. The consequent sinking of warm, subtropical waters could trigger a chain of events leading to the dramatic LPTM environmental changes.

Methane Hydrates: Fuel for Late Paleocene Thermal Maximum Warming?

A key ingredient in LPTM climate change was the pronounced warming of deep waters, which could initiate a variety of processes by positive-feedback mechanisms (e.g., Zachos et al., 1993). Massive hydrate dissociation caused by deep-water warming is the only mechanism yet identified that can provide a volume of carbon sufficient to explain the rate and magnitude of the global $\delta^{13}C$ shift (e.g., Dickens et al., 1995). If enough methane escaped the sediment column (e.g., Nisbet, 1990), it could have rapidly depleted oxygen in deep waters and fueled the LPTM warming through positive feedback (e.g., Dickens et al., 1997). The effect of hydrate carbon could have been augmented by increasing temperatures and lowering the CO_2 solubility of ocean waters. The combination of the circum-Caribbean eruptions, North Atlantic igneous province volcanism, and the release of massive quantities of hydrate carbon can explain the climatic and isotopic shifts associated with the LPTM.

CONCLUSIONS

Records of the late Paleocene thermal maximum from the Caribbean show evidence of dramatic environmental changes including warming of tropical surface waters, deep-water oxygen deficiency, a prominent extinction of deep-sea fauna, and reorganization of the global carbon cycle. Multiple volcanic tephra layers, including one laid down at the onset of the warming event, suggest that an intense episode of explosive volcanism influenced climate. We postulate that effusive volcanism in the North Atlantic led to long-term warming concentrated at high latitudes prior to the LPTM. The Caribbean eruptions slowed the rate of warming preferentially at low latitudes, further reducing the difference between high- and low-latitude SSTs, ultimately reversing the relative densities of these water masses, and shifting deep-water sources to the subtropics. The consequent deep-water warming led to widespread dissociation of methane hydrates that fueled further warming.

ACKNOWLEDGMENTS

Support provided by JOI-USSAC to Bralower and National Science Foundation grants EAR-94-05784 to Bralower, EAR-94-06099 to Zachos and Thomas, and EAR-96-29074 to Whitney. We thank K. Bice, C. Kelly, and L. Sloan for discussions, G. Koehler for technical assistance, and M. Leckie and A. Mix for reviews.

REFERENCES CITED

Anderson, T. F., and Arthur, M. A., 1983, Stable isotopes of oxygen and carbon and their application to sedimentologic and paleoenvironmental problems, *in* Arthur, M. A., and Anderson, T. F., eds., Stable isotopes in sedimentary geology: Society of Economic Paleontologists and Mineralogists Short-Course 10, p. 1.1–1.151.

Bice, K. L., 1997, An investigation of early Eocene deep water warmth using uncoupled atmosphere and ocean general circulation models [Ph. D. thesis]: State College, Pennsylvania State University, 363 p.

Dickens, G. R., O'Neil, J. R., Rea, D. K., and Owen, R. M., 1995, Dissociation of oceanic methane hydrate as a cause of the carbon isotope excursion at the end of the Paleocene: Paleoceanography, v. 10, p. 965–971.

Dickens, G. R., Castillo, M. M., and Walker, J. G. C., 1997, A blast of gas in the latest Paleocene: Simulating first-order effects of massive dissociation of oceanic methane hydrate: Geology, v. 25, p. 259–262.

Eldholm, O., and Thomas, E., 1993, Environmental impact of volcanic margin formation: Earth and Planetary Science Letters, v. 117, p. 319–329.

Gagan, M. K., and Chivas, A. R., 1995, Oxygen isotopes in western Australian coral reveal Pinatubo aerosol-induced cooling in the Western Pacific Warm Pool: Geophysical Research Letters, v. 22, p. 1069–1072.

Hirschmann, M. M., Renne, P. R., and McBirney, A. R., 1997, $^{40}Ar/^{39}Ar$ dating of the Skaergaard intrusion: Earth and Planetary Science Letters, v. 146, p. 645–658.

Kaiho, K., Arinobu, T., Ishiwatari, R., Morgans, H., Okada, H., Takeda, N., Tazaki, N., Zhou, G., Kajiwara, Y., Matsumoto, R., Hirai, A., Niitsuma, N., and Wada, H., 1996, Latest Paleocene benthic foraminiferal extinction and environmental changes at Tawanui, New Zealand: Paleoceanography, v. 11, p. 447–465.

Kennett, J. P., and Stott, L. D., 1991, Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Palaeocene: Nature, v. 353, p. 225–229.

Koch, P. L., Zachos, J. C., and Gingerich, P. D., 1992, Coupled isotopic change in marine and continental carbon reservoirs at the Palaeocene/Eocene boundary: Nature, v. 358, p. 319–322.

Minnis, P., Harrison, E. F., Stowe, L. L., Gibson, G. G., Denn, D. R., Doelling, D. R., and Smith, W. L., Jr., 1993, Radiative climatic forcing by the Mount Pinatubo eruption: Science, v. 259, p. 1411–1415.

Nisbet, E. G., 1990, The end of the ice age: Canadian Journal of Earth Sciences, v. 27, p. 148–157.

O'Connell, S., Chandler, M. A., and Ruedy, R., 1996, Implications for the creation of warm saline deep water: Late Paleocene reconstructions and global climate model simulations, Geological Society of America Bulletin, v. 108, p. 270–284.

Owen, R. M., and Rea, D. K., 1985, Sea–floor hydrothermal activity links climate to tectonics: The Eocene carbon dioxide greenhouse: Science, v. 227, p. 166–169.

Pindell, J. L., and Barrett, S., 1990, Geological evolution of the Caribbean region: A plate tectonic perspective: *in* Dengo, G., and Case J., eds., The Caribbean region: Boulder, Colorado, Geological Society of America, Geology of North America, v. H, p. 405–432.

Rampino, M. R., 1991, Volcanism, climatic change and the geologic record: Society of Economic Paleontologists and Mineralogists Special Publication 45, p. 9–18.

Rampino, M. R., and Self, S., 1992, Volcanic winter and accelerated glaciation following the Toba super-eruption: Nature, v. 359, p. 50–52.

Rea, D. K., Zachos, J. C., Owen, R. M., and Gingerich, P. D., 1990, Global change at the Paleocene/Eocene boundary: Climatic and evolutionary consequences of tectonic events: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 79, p. 117–128.

Schrag, D. P., DePaolo, D. J., and Richter, F. M., 1995, Reconstructing past sea surface temperatures: Correcting for diagenesis of bulk marine carbonate: Geochimica et Cosmochimica Acta, v. 59, p. 2265–2278.

Sigurdsson, H., Leckie, R. M., and Acton, G., 1997, Proceedings of the Ocean Drilling Program, Initial reports, Volume 165: College Station, Texas, Ocean Drilling Program, 865 p.

Sloan, L. C., and Thomas, E., 1997, Global climate of the late Paleocene: Modeling the circumstances associated with a climate "event," *in* Aubry, M.-P., Lucas, S., and Berggren, W. A., eds., Late Paleocene and early Eocene climatic and biotic evolution: New York, Columbia University Press (in press).

Storey, M., Duncan, R. A., Larsen, H. C., Pedersen, A. K., Waagstein, R., Larsen, L. M., Tegner, C., and Lesher, C. A., 1996, Impact and rapid flow of the Iceland plume beneath Greenland at 61 Ma: Eos (Transactions, American Geophysical Union), v. 77, p. 839.

Stott, L. D., 1992, Higher temperatures and lower pCO_2: A climate enigma at the end of the Paleocene epoch: Paleoceanography, v. 7, p. 395–404.

Thomas, E., and Shackleton, N. J., 1996, The latest Paleocene benthic foraminiferal extinction and stable isotope anomalies, *in* Knox, R. O., Corfield, R. M., and Dunay, R. E., eds., Correlation of the early Paleogene in Northwest Europe: Geological Society [London] Special Publication 101, p. 401–441.

Varekamp, J. C., Kreulen, R., Poorter, R. P. E., and Van Bergen, M. J., 1992, Carbon sources in arc volcanism, with implications for the carbon cycle: Terra Nova, v. 4, p. 363–373.

Zachos, J. C., Lohmann, K. C., Walker, J. C. G., and Wise, S. W., Jr., 1993, Abrupt climate change and transient climates during the Paleogene: A marine perspective: Journal of Geology, v. 101, p. 191–213.

Zielinski, G. A., Mayewski, P. A., Meeker, L. D., Whitlow, S., Twickler, M. S., and Taylor, K., 1996, Potential atmospheric impact of the Toba mega-eruption ~71,000 years ago: Geophysical Research Letters, v. 23, p. 837–840.

Manuscript received May 23, 1997
Revised manuscript received August 20, 1997
Manuscript accepted August 27, 1997