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Abrupt Climate Change and Transient Climates During the Paleogene: A Marine Perspective[1]

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ABSTRACT

Detailed investigations of high latitude sequences recently collected by the Ocean Drilling Program (ODP) indicate that periods of rapid climate change often culminated in brief transient climates, with more extreme conditions than subsequent long term climates. Two examples of such events have been identified in the Paleogene; the first in latest Paleocene time in the middle of a warming trend that began several million years earlier: the second in earliest Oligocene time near the end of a Middle Eocene to Late Oligocene global cooling trend. Superimposed on the earlier event was a sudden and extreme warming of both high latitude sea surface and deep ocean waters. Imbedded in the latter transition was an abrupt decline in high latitude temperatures and the brief appearance of a full size continental ice-sheet on Antarctica. In both cases the climate extremes were not stable, lasting for less than a few hundred thousand years, indicating a temporary or transient climate state. Geochemical and sedimentological evidence suggest that both Paleogene climate events were accompanied by reorganizations in ocean circulation, and major perturbations in marine productivity and the global carbon cycle. The Paleocene-Eocene thermal maximum was marked by reduced oceanic turnover and decreases in global $\delta^{13}C$ and in marine productivity, while the Early Oligocene glacial maximum was accompanied by intensification of deep ocean circulation and elevated $\delta^{13}C$ and productivity. It has been suggested that sudden changes in climate and/or ocean circulation might occur as a result of gradual forcing as certain physical thresholds are exceeded. We investigate the possibility that sudden reorganizations in ocean and/or atmosphere circulation during these abrupt transitions generated short-term positive feedbacks that briefly sustained these transient climatic states.

"The climate problem is as difficult as it is important. . ."

T. C. Chamberlin (1906)

Introduction

For decades, based on information derived primarily from studies of terrestrial and marine records on land, the Paleogene was viewed conservatively as a time of gradual climate transition between the warm, relatively ice-free Cretaceous, and the colder glacial Neogene. To a first approximation this perception of Paleogene climate is accurate; reconstructions of climate based on paleontologic proxies indicate that the Early Eocene was a time of exceptional warmth with near subtropical conditions in high latitude marine and terrestrial environments (Wolfe 1980; Estes and Hutchison 1980; Axelrod 1984; Francis 1988). By Late Eocene time, most of the Early Eocene flora and fauna had disappeared from high latitude regions as temperatures dropped and glacial conditions emerged (Wolfe 1992; Webb et al. 1984). With the possible exception of ice-sheet formation, which may have been episodic, it was generally assumed that the transition between these climate states was a relatively gradual process, occurring over millions of years.

This view of Paleogene climate, however, changed in the early 1970s as the Deep Sea Drilling Project (DSDP) began to recover continuous sedimentary sequences from the ocean floor. As more and more records were collected and examined, it became evident that the long term Paleogene climate transition was punctuated by a number of steps (Shackleton and Kennett 1975; Haq et al. 1977; Kennett 1977; Savin 1977; Keller 1983; Keigwin and Corliss 1986). While it was generally acknowledged that these steps represented important climatic transitions, details concerning the exact timing and magnitude of each remained unclear.

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This was due in part to disturbances and gaps in cores, which often deterred high resolution studies. With the inception of the Ocean Drilling Program and improved coring techniques in the mid 1980s, the quality of sedimentary sections improved. Also, exploration was conducted in regions previously considered inaccessible, most notably in the high latitude Southern Oceans.

High resolution isotopic studies of these new deep sea sequences immediately began to reveal details of Paleogene climate transitions not obvious in previous reconstructions. Imbedded within the long term climate signals were several short-term, dramatic climatic excursions that occurred over thousands of years. The first of these climatic events, which occurred near the Paleocene/Eocene boundary, was recognized in a pelagic sequence from Maud Rise, in the Atlantic sector of the Southern Ocean (Kennett and Stott 1991). Oxygen isotope analysis of foraminifera revealed abrupt, but brief, warming of deep ocean and high latitude surface waters coincident with a major extinction of benthic foraminifera (Tjalsma and Lohmann 1983; Miller et al. 1987*b*; Thomas 1990). This abrupt event occurred within a well-documented longer term warming that began in late Paleocene and culminated in Early Eocene time (Savin 1977; Shackleton and Kennett 1975; Miller et al. 1987*a*). The second prominent climatic event of the Paleogene occurred just above the Eocene/Oligocene boundary. This event, characterized by sudden deep water cooling and large scale ice-sheet expansion, was superimposed on a long term global cooling (Shackleton and Kennett 1975; Poore and Matthews 1984; Miller et al. 1987*a*; Zachos et al. 1992*b*). Both events appear to be transient climates that were not in equilibrium with the long term gradual climate transitions within which they were imbedded. These transient climates have been interpreted by some as climatic overshoots because the timing, direction, and rate of change suggest climate states that have briefly passed equilibrium conditions (Kennett and Stott 1991).

The long term changes in Paleogene climate have generally been attributed to mantle related processes (i.e., changes in paleogeography, oceanic gateways, sea level related to expansion of mid-ocean ridge and/or oceanic plateau volumes, mantle or volcanically derived CO_2) (e.g., Frakes and Kemp 1972; Shackleton and Kennett 1975; Berggren and Hollister 1977; Kennett 1977; Fisher and Arthur 1977; Schnitker 1980; Owen and Rea 1985; Barron 1985; Lasaga et al. 1985; Wise et al. 1985; McGowran 1989, 1990; Arthur et al. 1991). If so, what was the origin of these two abrupt Paleogene climate events? In the past, such non-Milankovitch-related, short-term climatic aberrations have been attributed to catastrophic events, such as meteorite impacts or prolonged volcanic eruptions (e.g., Bray 1979; Flohn 1979; Ganapthy 1982; Rampino and Stothers 1984). However, unlike the Cretaceous/Tertiary boundary, there is at present little direct or indirect evidence of such catastrophes coincident with these transitions (i.e., indiscriminate mass extinction, iridium anomalies, widespread microtektite or ash layers, etc.). Instead, it appears that these abrupt events were in some way integrally related to the longer term climate transitions upon which they were superimposed.

From examination of paleoclimate reconstructions and climate model simulations over the last decade, climatologists and geologists have begun to recognize that even under apparent gradual forcing, climate transitions may occur abruptly (e.g., Kennett and Shackleton 1976; Berger et al. 1981; Vincent and Berger 1985; Berger and Labeyrie 1987 [refs. therein]; Crowley and North 1988; Kennett and Stott 1991). Such phenomena have been attributed to the existence of threshold levels in complex physical systems where more than one equilibrium state is possible (e.g., Rooth 1982; Crowley and North 1988; Manabe and Stouffer 1988). Under gradual forcing, as thresholds are neared, slight additional forcing can result in sudden jumps between equilibrium climatic states. This concept provides an attractive explanation for why as boundary conditions have slowly changed over earth history, the climate has responded in a stepwise fashion. What is more inexplicable, however, is that some climate transitions were not only abrupt, but also produced short-term non-equilibrium climates that were more extreme than the earlier or subsequent climates. The existence of such transient climates is significant, given the lack of evidence for external forcing, because it highlights the potential of internal feedbacks to amplify or dampen global climate change.

Feedbacks involving changes in ocean circulation and chemistry are generally considered to be among the most viable mechanisms of generating non-linear responses during climate transitions (Brass et al. 1982*a*; Berger 1982; Broecker et al. 1985; Broecker and Denton 1990). Today's ocean circulation is thermohaline, driven mainly by differences in temperature. In highly frigid polar regions, the rate of sinking (bouyancy flux) is of sufficient magnitude to supply the entire deep ocean with cold, highly oxygenated water. However, the balance between temperature and salinity is such

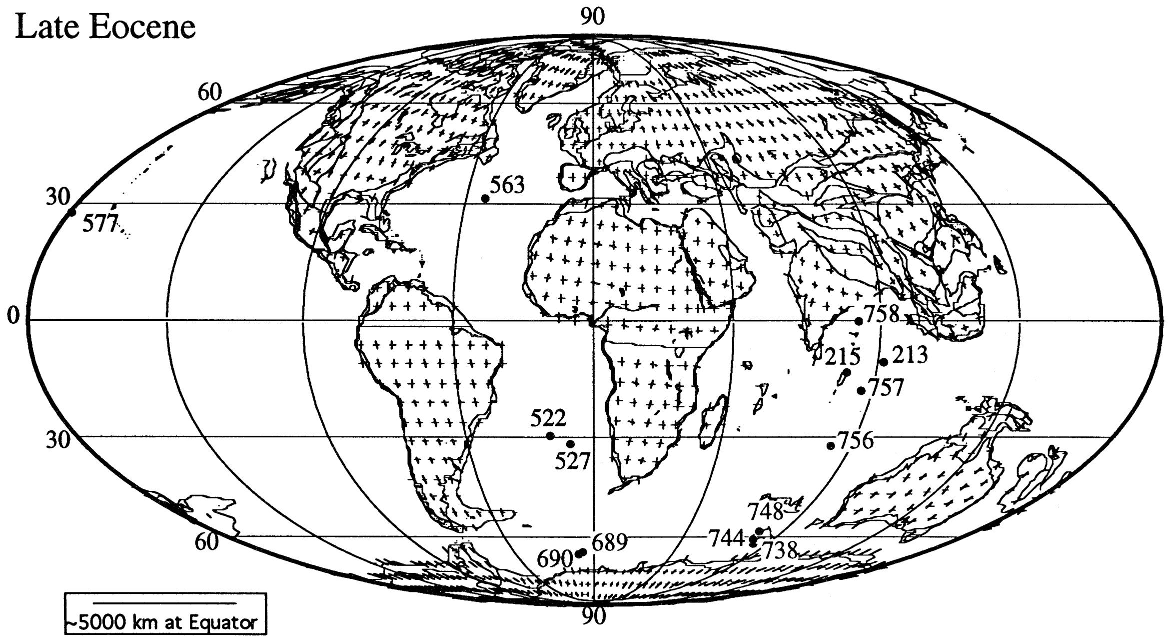

Figure 1. Paleogeographic reconstruction showing the late Eocene locations of DSDP and ODP sites considered in this study (map designed with Paleogeographic Information System™, M. Ross).

that during warmer times when planetary temperature gradients were lower, circulation might have been halothermal, that is driven more by salinity differences with considerable bottom water formation in low latitude, highly evaporative, marginal seas (e.g., Chamberlin 1906; Brass et al. 1982*a*, 1982*b*; Kennett and Stott 1990). The rate of deep water formation by this process, however, is more limited by salt availability and evaporation rates. As a result, halothermal circulation should be comparatively more sluggish than thermohaline circulation. The critical factor here, as first recognized by Chamberlin (1906), is that the overall balance between salinity and temperature on seawater density is such that only slight changes in the planetary temperature gradient might be necessary to push circulation from one mode to another. Because of the dominant role of the ocean in carbon cycling, sudden changes in the mode of circulation and deep sea ventilation rates can potentially affect the amount of carbon in the atmosphere (e.g., Brass et al. 1982*a*; Berger 1982; Broecker et al. 1985). For the two short-term climate transitions discussed above, considerable evidence exists to suggest that the modes and rates of ocean/atmosphere circulation changed. Moreover, it appears that each of these climatic transitions was accompanied by a significant change in the overall content and distribution of carbon within the ocean/atmosphere system. In light of this evidence, one must wonder if feedbacks involving ocean circulation and the carbon cycle served either to trigger or modify these abrupt climatic transitions.

In the following pages, we examine published evidence of gradual and abrupt climatic change, and transient climates during the Paleogene focusing on two specific episodes, the Late Paleocene-Early Eocene warming, and the Late Eocene-Early Oligocene cooling and glaciation. Paleogene isotopic records from Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) sites (figure 1) that bracket either one of these climatic transitions have been compiled. Included in this compilation are several new records from sequences recently recovered from the climatically sensitive, high latitude Southern Oceans. Oxygen isotope records are used to constrain the timing and scale of temperature and ice-volume changes through the Paleogene, while carbon isotopes are used to help identify shifts in ocean carbon content and distribution. Our primary objective here is to evaluate the evidence of abrupt climatic transitions and explore the prospect that changes in ocean/atmosphere circulation and chemistry may have triggered or amplified these events.

Paleogene Deep Sea Stable Isotope Records

Stable isotopes are among the most effective tools available for reconstructing prehistorical climate, not only for quantifying changes in temperature and ice-volume, but also as geochemical tracers

Table 1.

Site	Benthic Taxon	Source	Paleodepth & latitude at (35 Ma) or [55 Ma]	Primary Age Control
213	*N. truempyi*	a, l	[3500m, 30°S][a,o]	Nannofossil
215	*N. truempyi*	a, l, b	[3500m, 27°S][a,o]	Nannofossil
522	*Cibicidoides*	c	[700m, 36°S][m]	Magnet.
527	*Nuttallides*	d	[3400m, 37°S][p]	Magnet. & Nannofossil
563	*Cibicidoides*	e	(2200m, 32°N)[m]	Nannofossil
577	*N. truempyi, Aragonia*	f, g	[1900m, 20°N][f]	Magnet.
689	*Cibicidoides*	h	(1650m, 64°S)[h]	Magnet.
690	*Cibicidoides*	h	[2100m, 65°S][h]	Magnet. & Nannofossil*
738	*Cibicidoides & N. truempyi*	i	[1800m, 61°S][h,o]	Nannofossil
744	*Cibicidoides*	i	(1200m, 62°S)[h,o]	Magnet. & Nannofossil
748	*Cibicidoides*	j	(1100m, 59°S)[h,o]	Magnet. & Nannofossil
756	*Cibicidoides*	k	(500m, 43°S)[a,o]	Nannofossil
757	*Cibicidoides*	k	(1000m, 33°S)[a,o]	Nannofossil
758	*N. truempyi*	a, l	[2000m, 22°S][a,o]	Nannofossil

Sources. a, Zachos et al. 1992*c*; b, Hovan and Rea 1991; c, Miller et al. 1988; Shackleton et al. 1984; e, Miller and Thomas 1985; f, Miller et al. 1987*b*; g, Zachos et al. 1985; h, Kennett and Stott, 1990; i, Barrera and Huber 1991; j, Zachos et al. 1992*a*, 1992*b*; k, Rea et al. 1991; l, Zachos et al. unpub. data; m, Keigwin and Corliss 1986; n, Coffin 1992; o, Royer and Sandwell 1989, Royer et al. 1989; p, Moore et al. 1984.
* Original age model modified (see ref. a).

and as a means of improving stratigraphic correlations. In fact, our perception of Cenozoic climate has been influenced largely by compilations of benthic foraminiferal stable isotope records from deep sea sequences (e.g., Douglas and Savin 1975; Shackleton and Kennett 1975; Savin 1977; Miller et al. 1987*a*). Such benthic foraminifera isotope records, because of the physical and chemical homogeneity of the deep sea, are generally considered to be most representative of long-term global environmental change.

As a frame of reference for our discussion of Paleogene climate, we have constructed a Paleogene benthic foraminifer stable isotope record using published data from 14 ODP and DSDP sites (see table 1 for references) (figure 2). This compilation includes some of the data from the Miller et al. (1987*a*, 1987*b*) Atlantic and Pacific reconstructions as well as new data from Southern and Indian Ocean sites. The majority of data are of the genera *Cibicidoides* and *Nuttallides*. The Paleogene paleodepths cover a broad range, but generally fall between 1000 and 3500 m (table 1). All age models are calibrated to the geomagnetic polarity time scale of Berggren et al. (1985), and for future comparison, cross-calibrated to the revised time scale of Cande and Kent (1992).

Computations of Paleogene Deep Sea Temperatures and Ice-Volume. The benthic foraminiferal oxygen isotope record forms the basis for our calculations of Paleogene bottom temperatures and global ice-volumes (figure 3). Paleotemperatures are based on the equation of Erez and Luz (1983),

$$T°C = 16.998 + -4.52(\delta^{18}O_{CC} - \delta^{18}O_{SW}) + 0.028(\delta^{18}O_{CC} - \delta^{18}O_{SW})^2$$

where $\delta^{18}O_{CC}$ is the isotopic composition of the foraminifer $CaCO_3$ relative to PDB and $\delta^{18}O_{SW}$ is the isotopic composition of sea water ($\delta^{18}O_{seawater}$ (SMOW) -0.27) in which the calcite was precipitated. Comparisons show that, of the available temperature equations, this equation most closely approximates predicted equilibrium values in modern marine settings (Zahn and Mix 1991). To correct for foraminifer disequilibrium vital effects, the $\delta^{18}O$ values of *Cibicidoides* and *Nuttallides* were adjusted by $+0.6$ and $+0.4$, respectively (Shackleton et al. 1984).

Our calculation of global marine temperature from Late Paleocene to Middle Eocene is relatively straightforward, since an ice-free world could be assumed for this period ($\delta^{18}O_{SW}$(SMOW) = -0.93‰). Temperature calculations for intervals above the Middle Eocene, however, include the added affect of ice-volume changes as a possible source of $\delta^{18}O_{CC}$ variation.

While expeditions around Antarctica have produced glaciomarine sediments extending as far

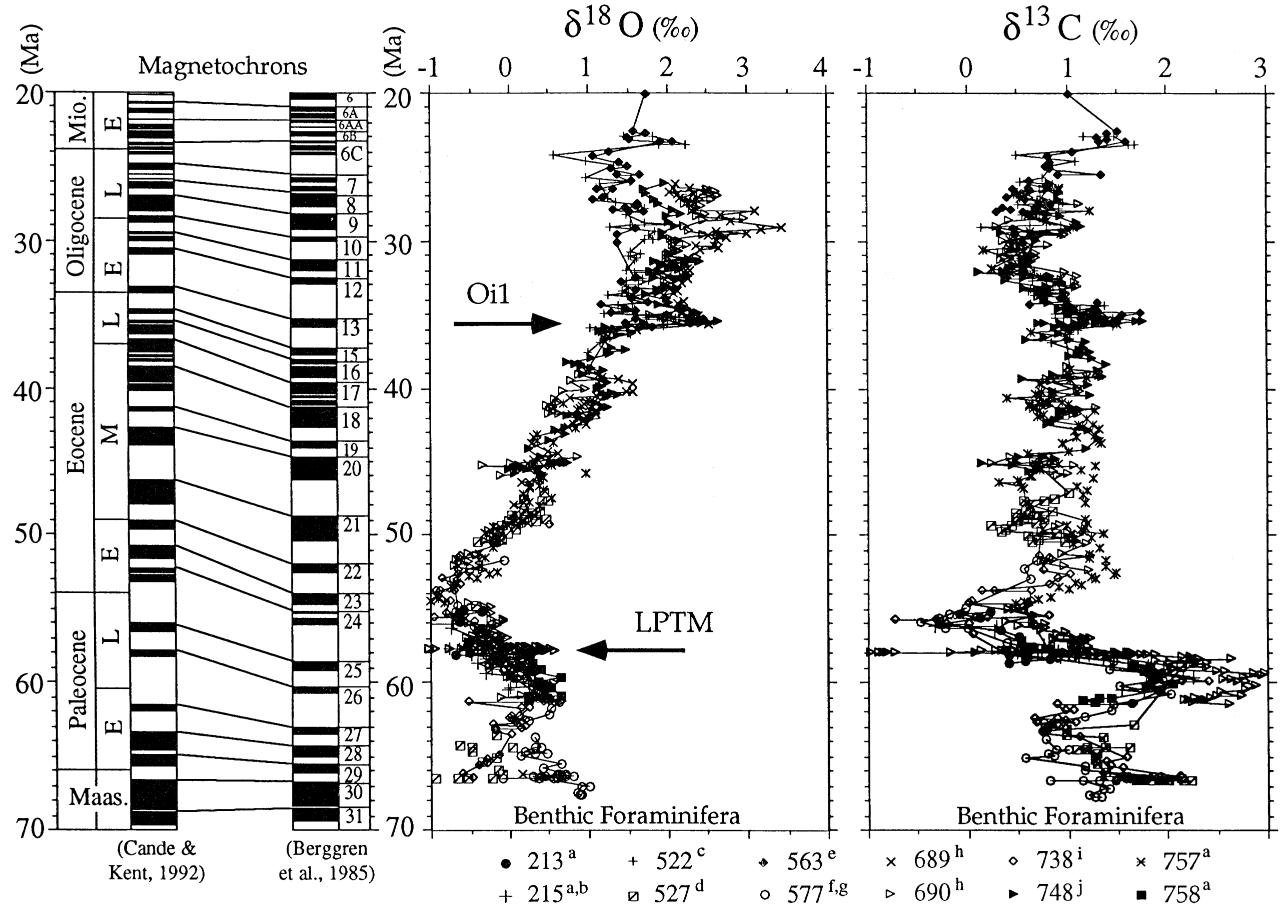

Figure 2. Compilation of benthic foraminifer carbon and oxygen isotope data from DSDP and ODP sites plotted versus age. Numerical ages are calibrated to the time scale of Berggren et al. (1985) and cross calibrated to the time scale of Cande and Kent (1992). Data represent measured values of *Cibicidoides* spp. or *Nuttallides* spp. (table 1) relative to the Pee Dee Belemnite Standard (PDB) (Sources: *a*, Zachos et al. 1992*c*; *b*, Hovan and Rea 1991; *c*, Miller et al. 1988; Shackleton et al. 1984; *e*, Miller and Thomas 1985; *f*, Miller et al. 1987*b*; *g*, Zachos et al. 1985; *h*, Kennett and Stott 1990, 1991; *i*, Barrera and Huber 1991; *j*, Zachos et al. 1992*a*, 1992*b*). Arrows show the approximate positions of the Late Paleocene Thermal Maximum (LPTM) and Early Oligocene Glacial Maximum (Oi1) events. Note that the numerical ages of the Paleocene-Eocene and Eocene-Oligocene boundaries have been revised to 55 and 34 Ma, respectively (Cande and Kent 1992).

back as Middle Eocene time (Margolis and Kennett 1975; Hayes et al. 1975; Leckie and Webb 1983; Webb et al. 1984; Barrett et al. 1989; Barron et al. 1989; Schlich et al. 1989), it appears that the earliest widespread glacial activity did not occur until the earliest Oligocene. This is clearly depicted in a graph showing the temporal distribution of Paleogene ice-rafted debris (IRD) in sequences from the shelves and marginal seas of Antarctica (figure 3) (Wise et al. 1991). In near shore settings, Oligocene glaciomarine sediments have been recovered from McMurdo Sound and Weddell Sea (Barrett et al. 1989, Barker et al. 1988), while sediments as old as Late Eocene have been found in Prydz Bay (Hambrey et al. 1991). Paleogene glaciomarine sediments have also been recovered in offshore settings several hundred kilometers away from Antarctica. At two ODP Sites, 744 and 748, on Kerguelen Plateau, traces of IRD were discovered in lower Oligocene pelagic sediments (Barron et al. 1989; Schlich et al. 1989). The IRD, which consisted of fractured quartz grains and heavy minerals as large as several mm in diameter, was apparently transported by icebergs from Prydz Bay to the south (Ehrmann 1991; Breza and Wise 1992; Ehrmann and Mackensen 1992).

These new records of IRD deposition facilitate interpretation of the oxygen isotope record in terms of temperature and ice-volume, because they provide more direct information on the timing and extent of glacial activity on Antarctica. Before they were available, certain assumptions were required

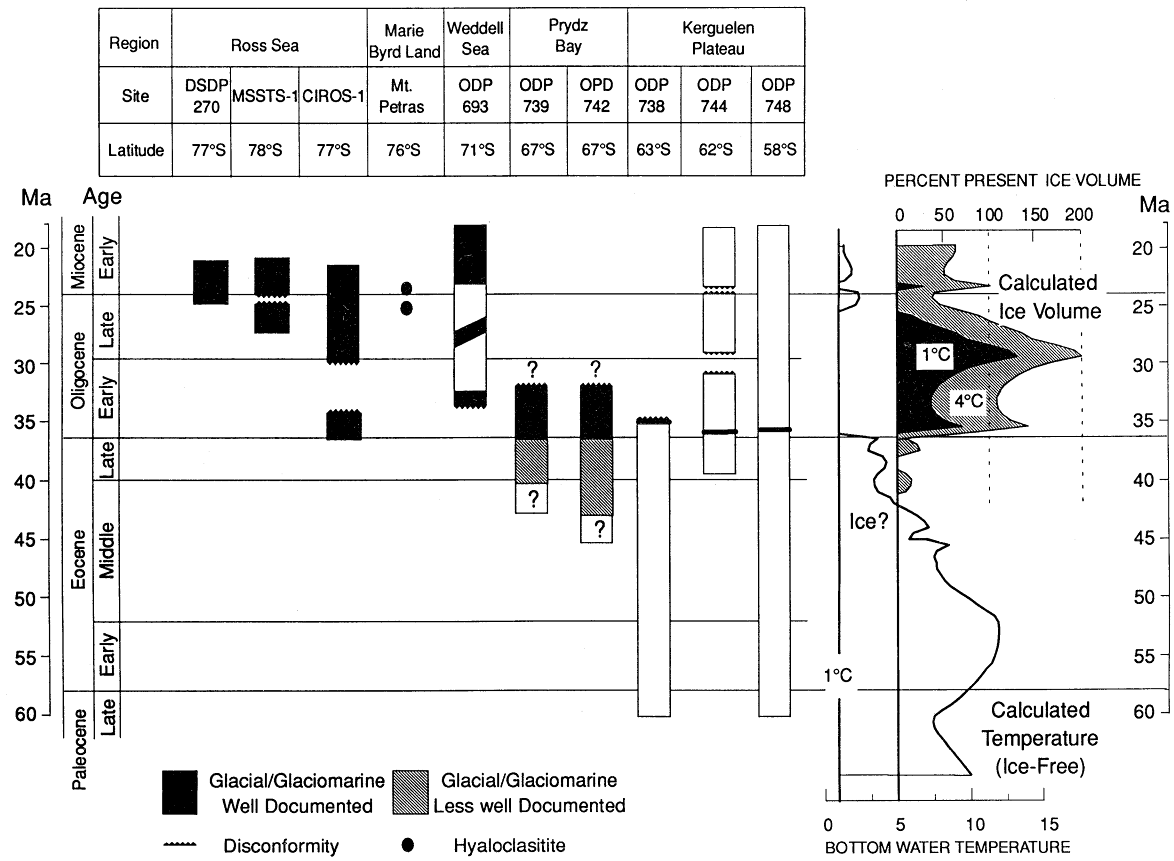

Figure 3. Distribution of confirmed Paleogene glaciomarine sediments in drill cores from Antarctic and Southern Ocean locations (Wise et al. 1991) plotted along with the record of deep sea temperatures and global ice-volume (as a % of present day ice-volume) as computed from the benthic foraminifer oxygen isotope record (figure 2). Isotopic values were adjusted (+0.6‰) for disequilibrium vital effects. The mean isotopic composition of continental ice was assumed to be constant at −45‰ (SMOW). Ice-volumes were calculated for deep water temperatures of 1 and 4°C. The 0% ice-volume line provides a lower limit on deep sea temperatures. The black shaded region is the lower limit on ice-volume assuming temperatures no colder than 1°C, the stippled region is the computed range of ice-volumes assuming deep water temperatures between 1 and 4°C. The numerical ages in this plot are based on the time scale of Berggren et al. (1985).

to deconvolve the ice-volume from the temperature signal in oxygen isotope records (Matthews and Poore 1981; Poore and Matthews 1984; Keigwin and Corliss 1986; Miller et al. 1987; Prentice and Matthews 1988). These assumptions typically involved eliminating temperature as a variable in the oxygen isotope temperature equation. By compiling records from locations where temperatures were either limited and/or should be constant, one could attribute all variability to ice-volume. Miller et al. (1987*a*), in considering benthic foraminifera $\delta^{18}O$ records from Atlantic DSDP sites, noted that if one assumed ice-free conditions for those intervals in which *Cibicidoides* values exceeded 1.8‰ (or 2.4‰ in equilibrium benthic values), bottom water temperatures as cold or colder than present day are required, a condition deemed to be unrealistic on a supposed ice-free earth. Thus, by invoking a lower limit for deep water temperatures, all variations involving values >1.8‰ could be attributed to ice-volume. While this method provided a lower limit on ice-volume, it could not constrain the upper limit; one could argue that substantial volumes of ice existed when values were lower by invoking higher deep water temperatures. As an alternative approach, Matthews and Poore (1981) and Prentice and Matthews (1988) supplemented benthic foraminiferal oxygen isotope records with planktonic records from western equatorial sites. By assuming constant sea surface temperatures (SST) in western equatorial regions, all temporal variability in the planktonic record was attributed to ice-volume change; variations in the difference between planktonic and benthic records were then interpreted as changes in deep ocean temperatures. In principle, this approach should provide more accurate constraints on ice-volume. In practice, however, there is a high potential for error due to dissolution effects. Studies of modern sediments have demonstrated that sea floor dissolution can significantly alter the oxygen isotopic composition of planktonic foraminifer assemblages by removing the more thin-walled, warm water, near-surface specimens, thereby shifting the mean $\delta^{18}O$ of the remaining population toward higher values (Erez 1979; Wu and Berger 1989).

Following the approach of Miller et al. (1987), we use the oxygen isotope record of benthic foraminifers and assume that deep water temperatures never fell below 1°C. This establishes a lower limit on ice-volume; when equilibrium $\delta^{18}O$ (PDB) values exceed 2.4‰, continental ice must have existed. In addition, we can characterize comparatively ice-free intervals by assuming that the known distribution of IRD accurately reflects historical glacial activity. As demonstrated above, the record from marine sequences indicates that little or no IRD was being deposited adjacent to Antarctica prior to the Late Eocene. Thus, we assume that either ice-sheets were absent or too small noticeably to affect the sea water $\delta^{18}O$ prior to Late Eocene time. We also assume that bottom waters formed mainly in the high southern latitudes over most of this time, an assumption supported by the fact that over much of the Paleogene (1) high latitude SST and deep sea temperatures covary, and (2) carbon isotope distributions suggest the youngest bottom waters were consistently near Antarctica (Pak and Miller 1992).

With these constraints, it is possible to provide a somewhat more quantitative interpretation of the Paleogene deep sea $\delta^{18}O$ record in terms of ice-volume and deep sea temperatures for the Paleogene (figure 3). Several important points emerge from this reconstruction. First, ice-sheets did not appear until earliest Late Eocene time when deep water temperatures—and we assume high latitude SST—dropped below 5–6°C, somewhat consistent with results of Community Climate Model (CCM) experiments that predict limited ice-accumulation on Antarctica when high latitude SST are substantially warmer than present day (Oglesby 1989). Second, the Oligocene $\delta^{18}O$ values indicate a permanent ice-sheet on Antarctica for the entire Oligocene. This agrees well with the somewhat incomplete records of IRD deposition from near-shore regions (figure 3), and the more complete records of clay mineral deposition in offshore pelagic regions (Maud Rise and Kerguelen Plateau), which show a permanent shift from smectite to illite/chlorite dominated assemblages beginning in Early Oligocene time (Ehrmann and Mackensen 1992). Third, at the peak of the Early Oligocene oxygen isotope event, global ice-volume must have exceeded 70% of present day ice-volume. Ice volume might have been greater if either, (1) deep water temperatures were warmer than 1 to 2°C, or (2) continental ice was more enriched in O^{18} than modern ice-sheets (e.g., Prentice and Matthews 1988). Regardless, an ice-sheet with at least 70% the mass of present day continental ice sheets is required, consistent with the widespread distribution of lower Oligocene glaciomarine sediments. Fourth, between the first appearance of IRD in the late Eocene, and full scale continental glaciation in the Early Oligocene, ice-sheets existed but of considerably lesser volume than present. While in principle one could argue for alternative interpretations of the Paleogene oxygen isotope record (i.e., large ice-sheets throughout the Paleogene) (Prentice and Matthews 1988), the assumptions and interpretations provided here are most consistent with independent paleontologic and sedimentologic proxies and theoretical considerations of climate.

Latest Paleocene Thermal Maximum (LPTM). The first major climatic event of the Paleogene occurred near the Paleocene/Eocene boundary, in the middle of a long term warming trend. The benthic foraminiferal oxygen isotope record shows that from the Late Paleocene to Early Eocene, deep sea temperatures warmed from 8° to 12°C (figure 3), during a time when low latitude temperatures remained more or less constant at levels near those of the present day (Shackleton and Boersma 1981; Boersma et al. 1987; Zachos et al. unpub. manuscript). The warm Early Eocene conditions were sustained for a period of about 2 to 3 myr, after which time deep waters began to cool. This long term record of deep sea temperatures bears a strong resemblance to paleontologic and isotopic records of high latitude sea surface temperatures for the same period. From Late Paleocene to Early Eocene time, high latitude planktonic foraminifer faunas and nannofossil floras became increasingly more diverse. Characteristic subtropical fauna and flora appeared in the Southern Oceans by latest Paleocene time, reached peak abundances in the Early Eocene, and slowly disappeared by the Middle Eocene (Haq et al. 1977; Pospichal and Wise 1990; Stott et al. 1990; Huber 1991). Also, planktonic foraminifer oxygen isotope records from high southern latitude sites tend to parallel the deep sea record for most of the Early Paleogene, indicating

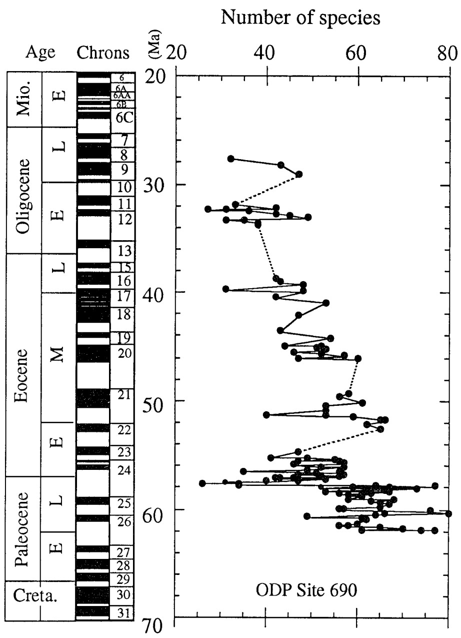

Figure 4. Variations in the total number of benthic foraminifer species present per sample at ODP Site 690, Maud Rise, plotted versus age (after Thomas 1991). The age model is based on the magnetostratigraphic interpretation of Spieß (1990) as calibrated to the time scale of Berggren et al. (1985). This plot illustrates that the largest change in Cenozoic species diversity occurred just below the Paleocene-Eocene boundary.

that the thermal evolution of these two regions were physically linked (Stott et al. 1990; Barrera and Huber 1991).

Perhaps the most striking evidence for environmental change, however, comes from benthic foraminifer species abundance records. Opposite of the patterns observed in planktic biota, diversity of deep-sea benthic foraminifera declined as epifauna taxa disappeared from the Late Paleocene to Early Eocene (figure 4) (Tjalsma and Lohmann 1983; Thomas 1990). Moreover, the turnover in benthic species composition was abrupt, with the majority of disappearances occurring just below the Paleocene/Eocene boundary (figure 4) (Miller et al. 1987*b*; Thomas 1990). Decreased productivity, deep water warming, and lower dissolved oxygen content have all been cited as possible causes of this sudden benthic crisis (Miller et al. 1987*b*; Thomas 1990; Pak and Miller 1992).

At Site 690B, Maud Rise in the Southern Ocean, this benthic foraminifer extinction horizon was located at the base of a 40 cm interval of laminated sediments (Thomas 1990). A high resolution isotopic study across this interval uncovered an unusual and dramatic short-term paleoenvironmental phenomenon (Kennett and Stott 1991). Imbedded within the long term warming at a level just below the Paleocene/Eocene boundary and synchronous with the benthic foraminiferal extinction horizon was an abrupt, short-term warming event (figure 5), an event we refer to here as the lastest Paleocene thermal maximum (LPTM). the oxygen isotope record of planktonic foraminifera shows that surface temperatures increased by 5 to 6°C during this event with maximum temperatures surpassing 20°C. The deep ocean warmed as well, by about 4°C, with peak temperatures in excess of 15°C. The entire transition to warmer temperatures appears to have taken place in less than 10 kyr. This thermal maximum, however, did not persist; almost immediately following the excursion, temperatures began to cool gradually over the next 100 kyr. Conditions eventually stabilized about 150 kyr later, but at levels higher than prior to the event.

An abrupt carbon isotope excursion was also found synchronous with the thermal excursion (figure 5). Benthic foraminifera $\delta^{13}C$ decreased by >3.0‰ in less than 10 kyr, while planktonic foraminifer values decreased by an even larger margin so that by the apex of this excursion, the $\delta^{13}C$ difference between planktonic and benthics was close to zero, implying that the surface to deep water carbon isotope gradient at this one site had completely disappeared. As observed in the temperature record, the $\delta^{13}C$ ratios gradually recovered, but to values about 1.0‰ lower than those prior to the event. In addition to the isotopic excursions, a decrease in bulk carbonate content has been recognized in this interval (figure 5) (O'Connell 1990).

As a result of the Site 690B findings, several investigations were immediately undertaken to locate this event at other pelagic sequences (Stott 1992; Pak and Miller 1992; Corfield and Cartlidge 1992). To date, these and other investigations have identified similar isotope excursions at the benthic extinction horizon in at least five other Paleocene/Eocene pelagic sequences. In addition, both the long and short-term carbon isotope excursions have been traced to at least one terrestrial section in North America (Koch et al. 1992). Perhaps, the most significant of these additional discoveries, was that of Stott (1992) at DSDP Site 47.2 on the Shatsky Rise in the northwest Pacific. During the Late Paleocene, Site 47.2 was located beneath subtropical waters at roughly 15°N latitude. As at Site 690B in the southern ocean, a benthic extinction was found synchronous with a carbon isotope excursion of about the same magnitude in latest Paleocene time (figure 6). However, unlike the event

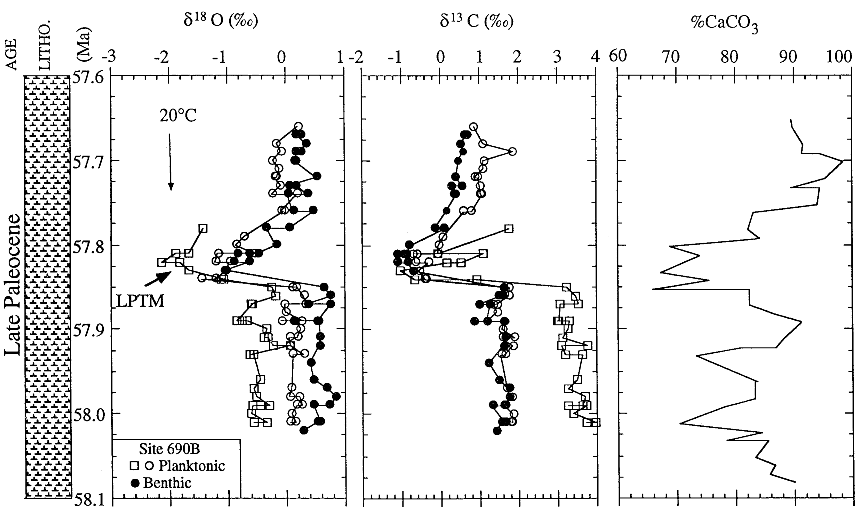

Figure 5. High resolution stable carbon and oxygen isotope record of planktonic and benthic foraminifers (Kennett and Stott 1991) and percent carbonate content (O'Connell 1990) from Core 690B-19H (167 to 174 mbsf). Numerical ages were recalculated using an alternative interpretation of the Site 690B magnetostratigraphy (see Spieß 1990; Zachos et al. 1992*c*) and the time scale of Berggren et al. (1985). As a result, the ages here are roughly 0.5 myr older than those of the original figure. Planktonic foraminifers include species of surface dweller *Acarinina* (open squares), and deep dweller *Subbotina* (open circles). Benthic foraminifers are *Nuttallides truempyi.* Graph shows an abrupt 5 to 6°C warming of surface and deep waters, accompanied by decreases in foraminifera $\delta^{13}C$, and bulk carbonate content.

at Maud Rise, planktonic foraminifer $\delta^{18}O$ values did not change at this tropical location, suggesting that as the deep oceans and high latitude surface waters rapidly warmed, low latitude surface temperatures remained more or less constant.

Early Oligocene Glacial Maximum (Oi1). The second major climatic event of the Paleogene occurred in the earliest Oligocene during a period of widespread IRD deposition around Antarctica. This event is marked by a prominent worldwide increase in foraminifera $\delta^{18}O$, superimposed on a longer term increase that began in the early Middle Eocene (figure 2). This longer term increase resulted from a 7 to 8°C cooling of deep sea waters (Kennett and Shackleton 1976; Savin 1977; Miller et al. 1987*a*). Details of the Early Oligocene $\delta^{18}O$ excursion, not obvious in the deep sea compilation, can be more clearly resolved in figure 7, where the oxygen and carbon isotope records of eight sites are plotted (Miller et al. 1987*a*, 1988; Kennett et al. 1990; Barrera and Huber 1991; Zachos et al. 1992*b*, 1992*c*). All $\delta^{18}O$ data in this comparison are of *Cibicidoides* adjusted by +0.6‰ to account for disequilibrium vital effects.

Several important relationships emerge from this comparison. First, despite given uncertainties in the individual age models, a $\delta^{18}O$ increase of variable magnitude was recorded near the period 35.6 to 35.8 Ma at every site. Most important is the oxygen isotope increase at Site 748 on Kerguelen Plateau, which coincides with a 40 cm IRD-bearing layer described above (figure 8) (Zachos et al. 1992*a*, 1992*b*). This record establishes a direct link between IRD deposition in the Southern Oceans and the ubiquitous Early Oligocene oxygen isotope increase. From here on, we will refer to the Early Oligocene glacial maximum as Oi1 following Miller et al. (1991).

Second, for sites with high sampling resolution it is clear that Oi1 was abrupt and brief, lasting for only a short period of time, 200 kyr at most, not taking into account the effects of sediment reworking due to bioturbation. The slight differences in the calculated duration of Oi1 between sites are probably artificial; linear interpolation of sedimentation rates between datum levels in age models prevents adjustment for minor hiatuses or short-term fluctuations in sedimentation. It also appears that at sites where the signal is not distinct, 756, 757, and 689, the spatial resolution of samples was inadequate (>1 sample per myr) to resolve the full isotopic expression of this short-term event.

Third, it appears that bottom temperatures during Oi1 were homogeneous, indicating that cold waters from the high latitudes briefly filled all ocean basins to at least 3500 m depth. However, in the interval of time following Oi1 until the end of the Oligocene, the distribution of $\delta^{18}O$ values between sites was much broader than in the time preceding this event (Barrera and Huber 1991;

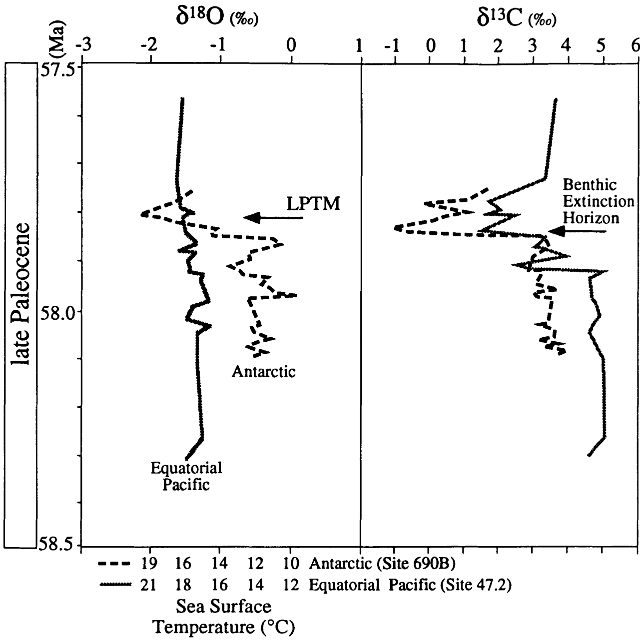

Figure 6. Comparison of latest Paleocene mixed layer planktonic foraminifer carbon and oxygen isotope values from tropical Pacific DSDP Site 47.2 (same location as Site 577) and Southern Ocean ODP Site 690B (figure 5) (Stott 1992). Numerical ages were recalculated based on an alternative interpretation of the Site 690B magnetostratigraphy (see Spieß 1990; Zachos et al. 1992*c*). The numerical ages are calibrated to the time scale of Berggren et al. (1985). This graph shows a brief 5 to 6°C warming of Antarctic surface waters during the late Paleocene during which tropical Pacific SST remain constant.

Zachos et al. 1992*a*). Benthic $\delta^{18}O$ values at Sites 522, 563, 747, were as much as 0.6‰ lower than those of sites 689, 690, 744, and 748. These intersite differences are not depth-related but more a function of latitude; the sites characterized by enriched values during the Oligocene were all located in the high southern latitudes proximal to Antarctica while the sites characterized by depleted values were located north of 40 to 50°S latitude (figure 1). This evidence further supports the idea that a permanent thermal re-configuration of deep and intermediate waters took place at this time (e.g., Murphy and Kennett 1986; Kennett and Barker 1990).

Finally, equilibrium $\delta^{18}O$ values at several sites exceed 3.0‰ at the peak of the Oi1 event (figure 7*a*), implying that a full scale ice-sheet(s) existed on the Antarctica continent, but only for a short period of time, an interpretation consistent with the evidence of brief, but widespread ice-rafting. The ice-sheet was most likely a wet-based, temperate ice sheet confined to east Antarctica (Ehrmann and Mackensen 1992). The temperature decrease associated with this ice-sheet expansion was probably less than a few degrees, a scenario compatible with the lack of major faunal turnover in pelagic benthic foraminifers at this time (e.g., Snyder et al. 1984; Corliss and Keigwin 1986; Thomas 1992). Following the brief expansion, a smaller ice-sheet probably remained until Middle Oligocene time, the next period of major expansion (Shackleton 1986; Miller et al. 1987; Kennett and Stott 1990).

Analogous to the LPTM event, the Oi1 climate event was accompanied by an abrupt carbon isotope excursion, but in the opposite direction. At each site, a substantial, but short-lived increase in $\delta^{13}C$ was recorded by benthic foraminifer coincident with the oxygen isotope increase at about 35.5 Ma ($\pm$0.2 myr) (figure 7*b*). The magnitude of increase recorded at sites plotted here varies from 0.4 to 0.8‰ with peak values ranging from 1.3 to 1.7‰. In the ensuing time from the Early to mid-Oligocene, benthic foraminifera $\delta^{13}C$ values at all sites declined by an average of 0.8‰. The peak earliest Oligocene values represent the highest recorded by benthic foraminifera between the Late Paleocene and Middle Miocene. This carbon isotope excursion does not appear to be confined to the deep ocean. At the few sites where planktonic foraminifer isotope records were constructed across the Eocene/Oligocene boundary, the planktonics show a similar if not slightly smaller increase in $\delta^{13}C$ (Vergnaud Grazzini and Oberhänsli 1986; Barrera and Huber 1991; Zachos et al. 1992*a*). This excursion, as that recorded during the latest Paleocene thermal maximum, has important implications for the possible origin of this Early Oligocene climatic extreme.

Discussion

The driving forces behind long term global climate change are still not well understood. The cooling of high latitudes during the Eocene is often attributed to changes in the configuration of continents and ocean gateways and associated effects on ocean heat transport (Kennett and Shackleton 1976; Berggren and Hollister 1977; Barker and Burrell 1977; Berggren 1982; Murphy and Kennett 1986; Bartek et al. 1992). Plate tectonic reconstructions show that, as Australia migrated away from Antarctica during the Eocene, ocean passages began to widen allowing for the eventual development of a shallow circumpolar current. In theory, this increased the thermal isolation of Antarctica by preventing warm currents and air masses from reaching the continent. Results of geochemical modeling exper-

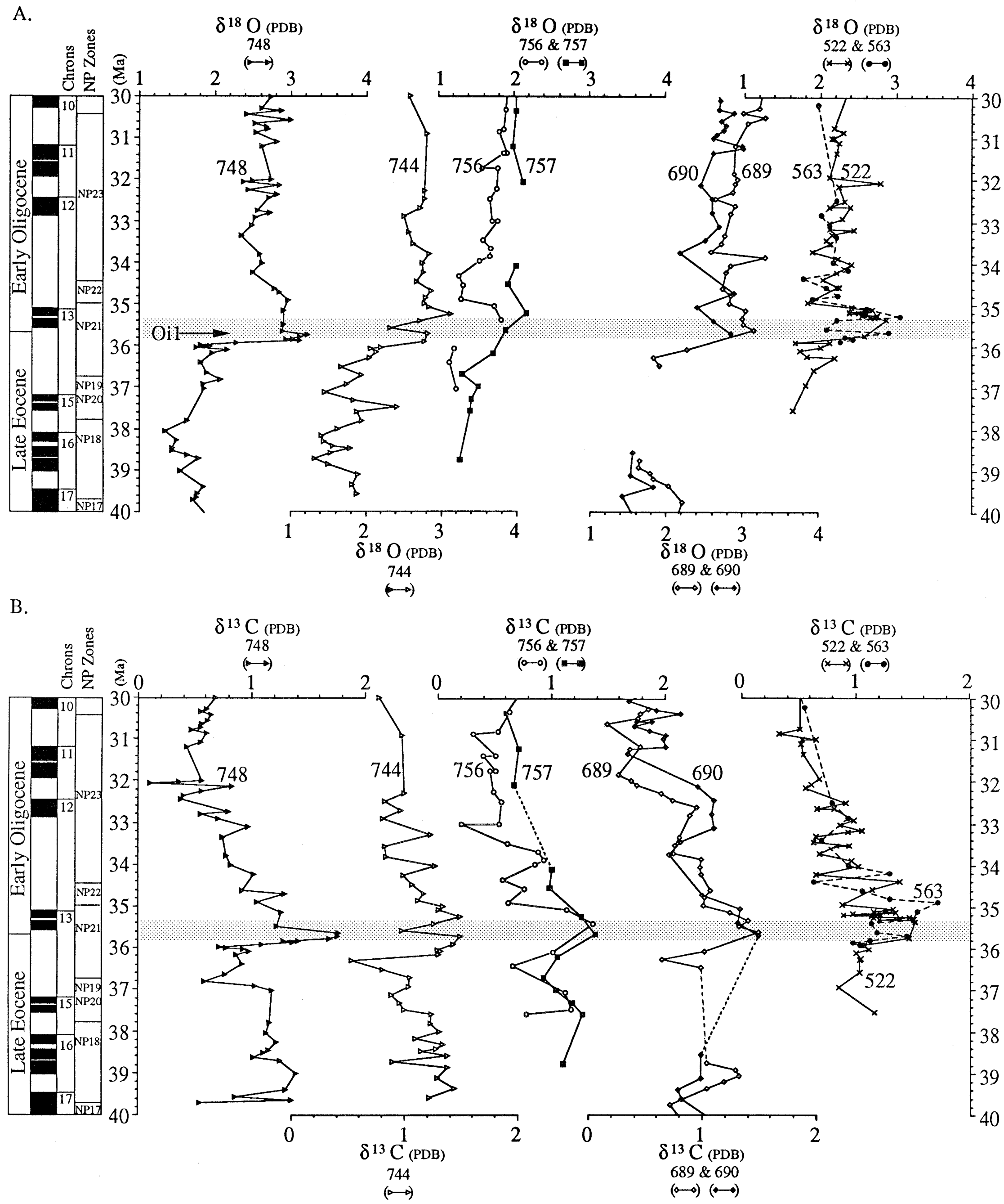

Figure 7. Benthic foraminifer *Cibicidoides* spp. (*a*) oxygen and (*b*) carbon isotope records from DSDP and ODP Sites 522 (Miller et al. 1988), 563 (Miller and Thomas 1985), 689, 690 (Kennett et al. 1990), 744 (Barrera and Huber 1991), 748 (Zachos et al. 1992*a*, 1992*b*), 756, and 757 (Rea et al. 1991). All values shown in this figure have been adjusted by +0.6‰ to account for disequilibrium vital effects. Age models were calibrated to the time scale of Berggren et al. (1985). The shaded region represents the interval of peak $\delta^{18}O$ and $\delta^{13}C$ values (Oi1).

iments, on the other hand, suggest that significant changes in the levels of greenhouse gasses should have occurred during the Paleogene (Lasaga et al. 1985; Berner 1991), an outcome which seems to be supported by studies employing geochemical proxies of pCO_2 (Arthur et al. 1991; Freeman and Hayes 1992). Preliminary results indicate that Early Eocene pCO_2 levels were nearly 6× present day. Other mechanisms for climate change (i.e., land-water distribution and albedo, sealevel, etc.) have been considered as well (e.g., Frakes and Kemp 1972; Barron et al. 1984). These paleogeographic

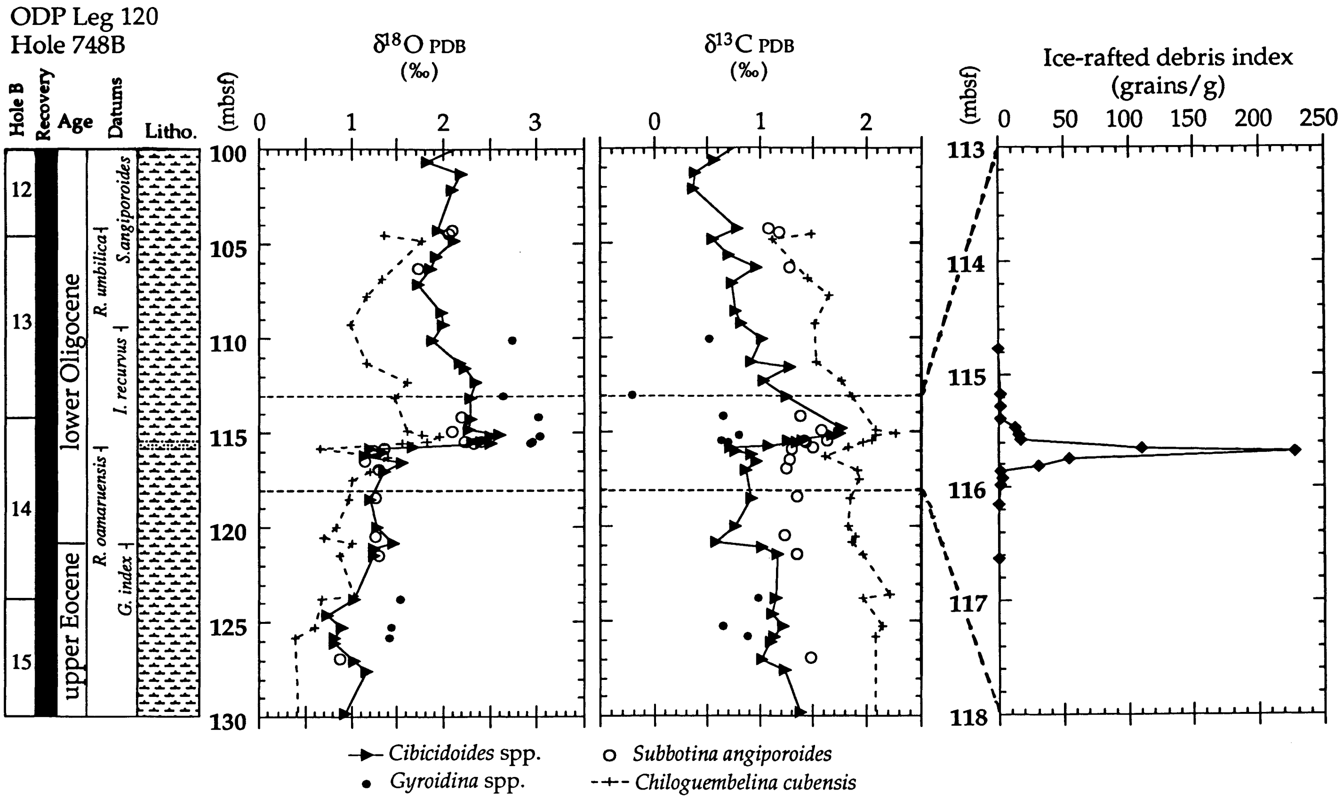

Figure 8. Site 748 upper Eocene to lower Oligocene stable carbon and oxygen isotope record of planktonic and benthic foraminifers from 100 to 130 m below the sea floor [bsf] (Zachos et al. 1992*b*) and ice-rafted debris concentrations (grains/g) from 113 to 118 mbsf (Breza and Wise 1992). Core recovery, planktonic foraminifer and calcareous nannofossil last occurrences, and lithology are in the left column. The peak in IRD abundance coincides precisely with the peak in benthic $\delta^{18}O$ in the lower Oligocene.

and geochemical forcing mechanisms may have acted independently or together, linked through tectonics. Regardless, because of the lack of convincing evidence for catastrophic external forcing (i.e., impacts, or volcanic eruptions) during the latest Paleocene thermal maximum or Early Oligocene glacial maximum, we are left to ponder how gradual forcing mechanisms might have created these abrupt, short-term climate aberrations.

One possible solution comes from climate theory and modeling. As noted by Crowley and North (1988), slowly changing boundary conditions in complex systems often lead to abrupt responses. This phenomenon has been attributed to the existence of threshold levels with multiple equilibrium states (Rooth 1982; Manabe and Stouffer 1988). When global climate reaches such a level, slight additional forcing can cause the system to move rapidly from one state to another. This model may apply to any aspect of the climate system. For example, an ice-sheet might have appeared on Antarctica only after high latitude summer temperatures fell below a critical threshold (Flohn 1979; Berger et al. 1981), or when winter precipitation exceeded a certain level (Frakes and Kemp 1972; Bartek et al. 1992), or for an already-existing small ice-cap, after it surpassed a critical size (North 1984).

However, while the theory of climatic thresholds might account for the abrupt appearance of a large Early Oligocene ice-sheet, or rapid Late Paleocene warming, it does not fully explain the brief nature of these climate extremes. It is clear that the climate conditions achieved in each of these two events were not stable. The extreme warmth of the high latitudes and deep ocean during the LPTM was replaced after less than 100 kyr by the cooler, but still equable Early Eocene climate, and the Oi1 full scale continental ice-sheet was replaced by a smaller, stable ice-sheet after a period of no more than a few hundred thousand years, implying that these climate states were not at equilibrium, at least not long term equilibrium. Given the brevity of these events, their unique positions, precisely at the transitions between equilibrium climates, and the lack of evidence for external forcing, one must consider the possibility that both the Late Paleocene thermal maximum and the full size Early Oligocene ice-sheet were transient climates, the products of ephemeral feedbacks triggered by the rapid transition between two different climatic states. Here we explore this possibility.

Climate Change, Ocean, and Atmospheric Circulation. Feedback mechanisms can be both physical and chemical in nature. Some feedbacks, such as higher albedo due to increased ice or cloud cover, or enhanced greenhouse warming due to increased water vapor, tend to have almost immediate repercussions, while others, involving ocean heat transport for example, may be delayed slightly ($\sim 10^2$ to 10^3 yr). Still other feedbacks, especially those involving the geochemistry of oceans or sedimentary reservoirs, can be delayed due to the slow response times of deep ocean turnover and its large scale chemical reactions. In considering this and the fact that deep ocean circulation appears to be sensitive to slight changes in the distribution of salinity and temperature (Berger 1982; Broecker and Denton 1990), it seems reasonable to suggest that the oceans are a potential source of non-linear feedbacks during climate transitions.

Clearly the climate of the Southern Oceans and Antarctica have an important influence on global ocean and atmospheric circulation (e.g., Oerlemans and van der Veen 1984). The ventilation rate of the deep ocean is dependent on the rate of deep water formation in the Antarctic. Cold persistent winds generated over the ice-sheets cool surface waters enhance seasonal sea-ice formation, thereby increasing the density of sea water and creating large scale deep convection and sinking. Moreover, the high pole-to-equator thermal gradient that results when a polar ice-sheet is present intensifies zonal atmospheric circulation (Flohn 1983). More vigorous atmospheric circulation, in turn, drives surface currents creating divergence and upwelling of nutrient-rich thermocline waters, resulting in narrow belts of high productivity, especially in the polar oceans.

Modeling experiments have suggested that slight changes in high latitude temperatures and/or the degree of seasonality can have a significant effect on the extent of sea-ice formation, intensity of meridional atmospheric circulation, and precipitation patterns, and thus on the location and rate of deep water formation (Hoeffert 1990; Barron and Peterson 1991). As such, it is likely that in the past, gradual changes in high latitude temperatures have probably initiated, at some threshold, ice-sheet formation and/or major reorganizations of surface and deep ocean circulation patterns. Here we assume that ocean circulation can operate in any one of a variety of configurations constrained by two extreme modes, thermohaline and halothermal circulation. The former mode is characteristic of a cold bi-polar glacial world with steep planetary temperature gradients and vigorous atmospheric and ocean circulation (e.g., similar to present day) while the latter is associated with a greenhouse world with a low planetary temperature gradient and more sluggish atmospheric and ocean circulation (e.g., Early Eocene).

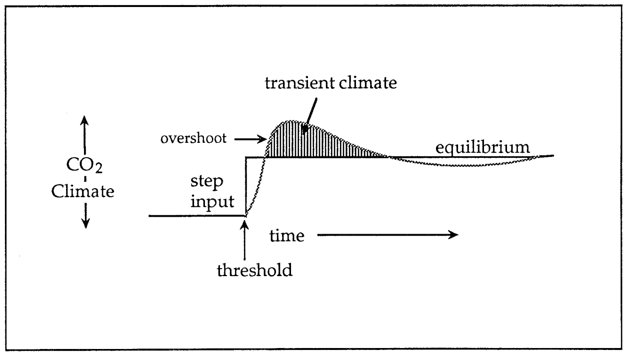

Figure 9. Schematic of expected non-linear response to a step input in the climate system (after Berger 1982).

An intermediate mode may have existed during the Late Paleocene when high latitude temperatures were warmer than today, but the bulk of deep waters still formed mainly in the southern oceans (Pak and Miller 1992). Because deep water temperatures were similar to those of high latitude surface waters at this time (Stott et al. 1990; Zachos et al. unpub. data), the contribution of warm-saline waters to the overall deep circulation was probably minor. However, as the high latitudes began to warm during the Late Paleocene and Early Eocene, the density distribution should have changed, shifting the ratio of deep water formation between the high and low latitudes. At some threshold, deep ocean circulation may have undergone a major transformation to a mode closer to halothermal circulation in character. While the exact complexion of this reorganization is uncertain, there is some isotopic evidence to suggest that the sources of deep water changed. According to Pak and Miller (1992), the distribution of benthic foraminiferal oxygen isotopes in the South Atlantic briefly changed during latest Paleocene time in response to what was an apparent incursion of intermediate to shallow deep waters from a non-high latitude source. In addition, there is considerable evidence to suggest that deep water turnover and ventilation rates slowed considerably. First, the extinction of epi-faunal benthic foraminifera is thought to have resulted from a drop in bottom water oxygen content (Thomas 1990). This interpretation is supported by other indicators of redox conditions; at one site, 690B on Maud Rise, the sediments over this interval are laminated indicating a brief cessation of benthic bioturbation, while at a number of Indian Ocean sites, bulk sediment manganese concentrations decrease dramatically across the Paleocene/Eocene boundary reflecting the emergence of sub-oxic conditions in pore waters (J. Dickens and R. Owen, pers. comm.). Such a reduction in bottom water oxygen is compatible with the sudden influx of warmer waters with low preformed oxygen contents.

Although no high resolution proxy records of atmospheric circulation are as yet available for this particular event, both theoretical and empirical evidence indicate that the overall low planetary temperature gradients of the Early Eocene resulted in less vigorous atmospheric circulation. Climate simulations with atmospheric circulation models using low planetary temperature gradients for the Eocene consistently produce lower global wind stress (Barron and Washington 1982; Sloan and Barron 1990), while reconstructions of Paleogene aeolian dust distributions in the Pacific and Indian Oceans show the smallest grain sizes in early Eocene time, implying less affective atmospheric transport (Rea et al. 1989; Hovan and Rea 1992). It seems, based on these findings, that additional warming of the high latitudes as is recorded during the LPTM would only serve further to reduce atmospheric wind intensity. A decline in the vigor of atmospheric circulation would dampen wind driven upwelling, augmenting the reduction in nutrient flux to the oceans photic zone.

In contrast, the isotope geochemistry and physical structure of Middle and Late Eocene sediments indicate a transition toward a more vigorous, thermohaline circulation (e.g., Kennett and Shackleton 1976; Corliss 1979; Keigwin 1980; Miller et al. 1985; Murphy and Kennett 1985). As the southern oceans cooled during the Middle and Late Eocene, it is likely that the rate of high latitude deep water formation increased (Kennett 1977; Kennett and Barker 1990). This transition appears to have proceeded in a step-wise fashion over the Eocene as successive thresholds were surpassed (Berger et al. 1981). Oxygen isotope records show a number of steps embedded within the long term transition (Kennett and Barker 1990; McGrowan 1990; Zachos et al. 1992*c*), each possibly reflecting some additional decrease in temperature. In terms of ice-volume, however, it appears that the largest step was that of the Oi1 event. As demonstrated here and elsewhere, this involved a >1.4‰ increase in $\delta^{18}O$, reflecting in large part expansion of a nearly full scale continental ice-sheet, as well as 2 to 3°C cooling of bottom waters. This was also when deep water formation rates and the vigor of bottom currents increased, as reflected by the increased distribution of Early Oligocene sedimentary hiatuses in southern ocean deeper water sequences (Kennett 1977; Keller et al. 1987; Ehrmann and Mackensen 1992). Hiatuses, however, were not restricted to the high southern latitudes; a widespread Early Oligocene reflector (horizon Au and reflector Ry) representing an erosional hiatus occurs throughout the deep North Atlantic. According to Miller and Tucholke (1983), this was created by a brief, but intense episode of North Atlantic deep water (NADW) production, an interpretation supported in part on reconstructed carbon isotope differences between the deep North Atlantic, Southern Ocean, and Pacific that show a pulse of NADW in the Southern Oceans during Oi1 (Miller et al. 1991). Rates of deep ocean turnover and ventilation would have increased substantially under this new configuration of widespread and voluminous deep water formation. With the exception of NADW production, it appears that reorganization of ocean circulation in Early Oligocene time was more or less permanent (e.g., Murphy and Kennett 1985; Zachos et al. 1992*b*).

In addition to increased deep water formation rates, it is probable that abrupt high latitude cooling and appearance of a full scale continental ice-sheet(s) on Antarctica in the Early Oligocene intensified zonal and regional atmospheric circulation. Today, the polar current is driven in part by strong persistent winds generated over the Antarctic ice-sheet. The boundary between the present day eastward-flowing polar current and westward-flowing circum-polar current is a zone of divergence and upwelling. This is important, since rates of net productivity along the Antarctic divergence are among the highest in the world. Thus, as high latitude climate cooled and ice-sheets began to expand in the Early Oligocene, one would expect an almost immediate increase in regional wind stress, and as a direct result, intensification of wind-driven upwelling.

Conceptual Model of Productivity and Carbon Cycle Response. The geochemical response of the oceans to sudden physical or chemical perturbation depends on many factors, including nutrient distributions, carbonate precipitation and dissolution, etc. So complex are the possible chemical interactions, that geochemical models are often required to test the various probable scenarios. Quantitative time-dependent modeling of transient responses, however, requires detailed knowledge of circulation rates, geochemical fluxes, and the sizes and distributions of reservoirs. Unfortunately, there is at present little understanding of even the most fundamental aspects of surface and deep ocean, and basin to basin communication of Paleogene oceans, such that the number of assumptions needed to model short-term transient responses would render most results highly ambiguous. Also, most box models typically lack the correct frequency response to forcing, which occurs at time scales less than the mixing time of the ocean reservoirs in such models (Southam and Peterson 1985). Rapid transient events become difficult to simulate, because feedbacks in the models tend to immediately dampen or reverse any shifts in ocean chemistry. As a result of these limitations, we provide only a conceptual model of the ocean-atmosphere carbon cycle in our discussion of potential geochemical responses of the Paleogene oceans to sudden reorganizations.

A better appreciation of the ocean's potential to modify climate through circulation changes can be gained by first considering the balance of readily exchangeable carbon on the earth's surface. The ocean contains 50× as much carbon as the atmosphere, most of which is stored in the deep ocean as dissolved inorganic carbon. Because of the relatively rapid gas exchange rate between the atmosphere and ocean, under steady state conditions the two are in near equilibrium. At steady state the concentration of CO_2 in the atmosphere is actually lower than it would be if it were at equilibrium with the whole ocean. This is a direct consequence of biological pumping of organic carbon (C-org) from the mixed layer to deep ocean (new production), which maintains a lower level of total CO_2 in the surface ocean mixed layer (Takahashi et al. 1981). The rate of organic carbon production in the mixed layer is controlled by nutrient availability, most of which is supplied from the subsurface ocean through vertical mixing and upwelling. Under steady state conditions with fixed rates of ocean turnover and productivity, a balance is maintained between the deep and surface layers and the surface layer and atmosphere. If, however, the rate of ocean circulation, upwelling, and/or riverine input of nutrients were to suddenly change, the rate at which nutrients are supplied to the mixed layer and the organic carbon flux to deep ocean would also change (Broecker 1982). Because the mixed layer and atmosphere equilibrate at a much faster rate than the surface and deep ocean (10s versus 1000s of years), the reservoirs could in theory become temporarily de-coupled, allowing for a transient excursion in atmospheric CO_2 (Southam and Peterson 1985). Furthermore, because organic carbon production rates should also affect the rate of carbon and/or nutrient burial, whole ocean TCO_2 could shift as well. Although carbonate precipitation and dissolution are capable of buffering these changes, it is possible for sizable lags to develop during rapid transitions. As a consequence, on short time scales atmospheric pCO_2 will be sensitive to slight changes in either (1) the mass of carbon in the ocean reservoir, (2) the gas exchange rate between surface and the atmosphere, or (3) the transfer of organic or inorganic carbon between the deep and surface ocean. For the moment we neglect the potential effects of the terrestrial biosphere while acknowledging its importance to the overall carbon balance.

The potential consequences of a sudden change in the rate of deep ocean circulation depend on several factors, mainly the direction and rate of change, but also where upwelling takes place (Toggweiler and Sarmiento 1985). In the Early Oligocene as high latitudes cooled and ice-sheets expanded, ocean and atmosphere circulation should have intensified. Productivity would increase as deep ocean turnover and wind-driven polar upwelling intensified. Elevated productivity, by transferring carbon to the deep ocean in particle flux would, in turn, draw down mixed layer and atmospheric CO_2 levels (Southam and Peterson 1985). Moreover, as particle flux increased, the amount sequestered into sediments should increase proportionately, resulting in net removal of carbon from the ocean. In addition to productivity effects, cooling of high latitude surface waters should increase CO_2 solubility in down-welling waters, adding slightly to the initial draw down of atmospheric CO_2 (e.g., Volk and Hoeffert 1985). An overall increase in pCO_2 could be prevented if the production and burial rate of biogenic carbonate increased proportionately with organic carbon. However, as shall be shown below, the Early Oligocene increase in organic carbon production was mainly in siliceous microfossils.

Conversely, as might have occurred during the LPTM, a sudden decrease in the vigor of ocean and atmosphere circulation due to high latitude warming and a reduction of the planetary temperature gradient should have just the opposite effect. Initially, as the rate of vertical mixing and deep water turnover decreased, nutrient fluxes would have declined, lowering the production and export of particulate carbon from the mixed layer. This in turn should lower the oceans capacity to take up CO_2 and even allow for slight degassing. This would be amplified by a warming induced decrease in CO_2 solubility in downwelling waters (Kennett and Stott 1991).

Late Paleocene and Early Oligocene Marine Paleoproductivity. Reduction or enhancement of marine productivity due to sudden changes in oceanic turnover rates and wind driven upwelling should leave unique and easily recognizable signatures in pelagic sediments. Of the various proxies of productivity, biogenic sediment accumulation rates are most widely used to evaluate changes in marine productivity. To a first approximation, the rate of biogenic sediment accumulation directly reflects rates of production in the surface ocean. Although this approach is somewhat limited by chronostratigraphic resolution and degree of preservation, it is most direct and least subject to interpretation. Also, the chemical or species composition of biogenic sediments can often serve as an important qualitative indicator of productivity. Biogenic opals are an important index of high productivity because they occur predominantly in regions where upwelling is intense and production extremely high. Thus, changes in the accumulation rate and character of sedimentation may be used for assessing regional changes in productivity. However, in order to extend this approach to evaluate global scale changes in organic carbon production and burial, one must also consider the carbon isotope record.

The content and distribution of stable carbon isotopes in the ocean are extremely sensitive to changes in organic carbon production, oxidation, and burial rates (see Keir 1991). While vertical carbon isotope gradients reflect in large part the rate of biological pumping of organic carbon from the surface to deep ocean (Kroopnick 1985), whole ocean $\delta^{13}C$ is sensitive to changes in the overall balance of carbon entering and leaving the ocean. Because organic matter carbon is highly depleted in C^{13} ($-20‰$), and carbon is cycled through the marine and terrestrial biospheres at relatively rapid rates, the production, oxidation, or burial of organic carbon strongly influence the distribution of carbon isotopes on the earth's surface. Given that nutrients are the limiting factor in marine primary production, and that the distribution of nutrients in the ocean is highly variable both in space and time, nutrient cycling is clearly one of the few processes capable of producing large changes in the ocean $\delta^{13}C$ over short periods of time ($\sim$1 to 100 kyr) (Kump 1990).

Several lines of evidence indicate that marine productivity diminished during the LPTM. First, at Site 690B where sediments were predominantly biogenic, carbonate contents dropped slightly (O'Connell 1990). This appears to reflect lower carbonate production and not increased dissolution, since foraminifer preservation did not noticeably change (Kennett and Stott 1991). Second, surface to deep water carbon isotope gradients as reconstructed from planktonic and benthic foraminifera briefly declined to near zero during the peak of the thermal maximum event (figure 5). Third, whole ocean $\delta^{13}C$ decreased by 2.5 to 3.0‰ during the warming episode (Kennett and Stott 1991; Pak and Miller 1992; Stott 1992). Such a rapid and large reduction in mean ocean $\delta^{13}C$ composition in less than 10 kyr requires an abrupt change in the flux of depleted carbon either to, or from the ocean. At the very least, it is consistent with a substantial reduction in C-org production and burial rate. Finally, as noted by Kennett and Stott (1991), with bottom water temperatures increasing to 15°C, preformed oxygen contents would have dropped sufficiently, and if productivity had remained constant, bottom waters should have become anoxic. There is no evidence from existing cores of total anoxia.

A brief reduction in marine productivity would immediately slow the uptake of CO_2 by the ocean which coupled with warming induced degassing of CO_2 (Brass et al. 1982; Kennett and Stott 1991) should have created a net export of CO_2 from the ocean to atmosphere, further amplifying global warming. Enhanced CO_2 levels would be temporary, however, as negative feedbacks would begin to reverse the process. For example, a slowly expanding oxygen minimum zone would elevate organic matter burial rates, or slowly rising nutrient levels might stimulate productivity. Also, warmer global temperatures would increase relative humidity and global precipitation rates (Barron and Peterson 1991), thereby stimulating terrestrial biomass production, as well as enhancing chemical weathering. In the process of lowering CO_2, each of the above-mentioned feedbacks would also tend to increase ocean/atmosphere $\delta^{13}C$.

In contrast, there is considerable evidence, both direct and indirect, that indicates that high latitude fertility increased during the earliest Oligocene Oi1 event. First, lithologies at Southern Ocean Sites 744 and 748 document a sudden but brief increase in the concentrations of diatoms and radiolaria coeval with the earliest Oligocene oxygen isotope increase (Schlich et al. 1989; Baldauf 1992). This short-term increase was apparently superimposed on a longer term, permanent rise in high latitude opal accumulation rates which had initiated a few million years earlier in latest Eocene time (Baldauf and Barron 1990). Second, the diversity of calcareous phytoplankton in the high latitudes dropped in apparent response to cooling, increased upwelling and eutrophication of surface waters (Haq et al. 1977; Aubry 1992). Third, the concentration of icthyolith remains also increased, reflecting the presence of highly fertile areas capable of supporting large populations of higher level organisms (Breza and Wise 1992). Finally, increased Southern Ocean productivity in the Early Oligocene is consistent with the simultaneous emergence of ancestral filter-feeding baleen whales in the high southern latitudes (Fordyce 1992). Large filter-feeding mammals would clearly benefit from the sudden appearance of narrow belts of wind driven upwelling.

Although the direct evidence of enhanced productivity is limited to the Southern Oceans, foraminifer carbon isotope records indicate that the effects on the carbon cycle were global. As described above, at every site, including DSDP Sites 574 and 77 from the Pacific (Miller and Thomas 1985), the Oi1 oxygen isotope increase was accompanied by a positive excursion in $\delta^{13}C$ of benthic foraminifera and, where analyzed, planktonic foraminifera (figure 7*b*). The excursion, more or less the same from 500 to 3200 m depth, ended with the highest $\delta^{13}C$ values in over 30 myr. Such a globally synchronous and uniform carbon shift most likely represents a change in the mean carbon composition of ocean DIC, not a temperature effect or redistribution of carbon isotopes between intermediate and deep reservoirs or from basin to basin (i.e., Vergnaud Grazzini and Oberhänsli 1986).

Given the evidence of circulation changes and enhanced productivity, it is not unreasonable to conjecture that the ubiquitous >0.7‰ increase recorded by benthic and planktonic foraminifera in the Early Oligocene was a by-product of a substantial increase in organic carbon burial rates that resulted from enhanced marine productivity (i.e., Vergnaud Grazzini and Oberhänsli 1986; Miller and Fairbanks 1985; McGowran 1989; Zachos et al. 1992*c*). This effect would have been amplified if much of the new production were concentrated within narrow polar and coastal upwelling zones. The increased production and burial of organic carbon would have created a short-term positive feedback on climate by lowering oceanic and atmospheric CO_2 levels, creating additional cooling. As with the late Paleocene event, however, this trend could not be sustained for long due to negative feedbacks. Elevated productivity by this mechanism alone would presumably persist for only several tens of thousands of years before nutrients concentrations adjusted to some steady state level, at which point, productivity and pCO_2 would stabilize, probably at a slightly higher level depending on the response of other feedbacks. This is assuming, of course, that the supply of nutrients to the ocean remained constant, a highly unlikely scenario considering the climatic and sea level changes brought about by this event. As sea level began to fall, large areas of continental shelf laden with nutrient rich sediments deposited over millions of years without interruption would have become immediately exposed and weathered, further stimulating productivity. In fact, to sustain high productivity levels for the full duration of Oi1, such additional nutrient inputs might be necessary. Similar ocean circulation/nutrient/productivity feedbacks have been postulated for the Middle Miocene (Monterey Hypothesis) (Vincent and Berger 1985) and, on a smaller scale, the last Glacial Maximum (LGM) (Broecker 1982).

In interpreting the carbon isotope and other records, we considered only a few of the processes that might affect the ocean/atmosphere carbon cycle. In the future, it will be necessary to consider other mechanisms that could rapidly perturb the carbon cycle. A sudden regression might change ocean $\delta^{13}C$ since oxidation of organic-rich shelf sediments would not only add nutrients to the ocean, but C^{13}-depleted carbon as well (Vincent and Berger 1985; Miller and Fairbanks 1985). Other mechanisms that might rapidly affect CO_2 and ocean $\delta^{13}C$ include the destruction and creation of terrestrial biomass (e.g., Shackleton 1977), changes in Redfield ratios of buried organic matter, or variations in the rate of continental weathering (Berger and Keir 1984; Vergnaud Grazzini and Oberhänsli 1986). Moreover, it appears the level of the calcite compensation depth dropped sometime near the Eocene/Oligocene boundary (van Andel 1975), a change that might tend to increase CO_2 levels depending on the overall mass balance of carbonate accumulation (Lasaga et al. 1985). Each of the above factors further complicates the already difficult process of estimating the effects of circulation and productivity changes on marine $\delta^{13}C$ and CO_2. Also, we have only touched upon sources of chemical feedbacks and their influence on greenhouse gasses. The importance of non-chemical feedbacks to climate transitions have been stressed in the past. For example, precipitation necessary for ice-sheet expansion in polar regions will also be sensitive to sudden changes in the temperature and large scale circulation of the ocean/atmosphere system (e.g., Schnitker 1980; Barron et al. 1989; Bartek et al. 1992). Nonetheless, whatever the source of feedback(s), given the abrupt and extreme nature of these climate transitions, it is clear that the feedback must be connected to a process capable of providing an abrupt, but temporary response. The production and burial of organic carbon in the ocean is one process with this potential. Because of the many complexities involved, the connection between high latitude climate, atmosphere, and ocean circulation and marine productivity during abrupt climatic transitions will eventually have to be examined with time-dependent numerical models constrained by proxies of the global carbon and nutrient cycles.

A final point to consider is the potential role of such abrupt events in the long-term evolution of biota. Recent improvements in the stratigraphic correlation of Paleogene marine and terrestrial sequences suggest that both these abrupt climatic excursions coincided with times of important evolutionary changes in terrestrial biota. Koch et al. (1992) presented carbon isotopic evidence from terrestrial sequences coupling precisely the LPTM event with the sudden emergence of several orders of mammals on land, including the early ancestors of all modern hoofed mammals (e.g., Gingerich 1989). They suggested that the short-term warming of high latitudes recorded in marine sections briefly expanded the biogeographic ranges of terrestrial fauna, previously confined to lower latitudes, into the vicinity of the northern land bridges connecting Eurasia and North America, temporarily permitting mass migration and mixing of mammals between continents. This hypothesis provides a plausible explanation for the sudden, but simultaneous appearance of these mammals in Eurasia and North America. Another prominent period in the evolution of terrestrial biota occurred 33 myr later in Early Oligocene time when European mammals were suddenly replaced by taxa from Asia and possibly North America (The Grande Coupre) (e.g., Hooker 1992). It has been suggested that this migration took place when interior seaways separating Asia and Europe withdrew due to a sudden glacio-eustatic lowering of sea level (Plint 1988). Recent revisions in the Eocene/Oligocene geochronology of terrestrial sequences indicate that this event took place very near the time of the Early Oligocene glacial maximum (Prothero and Swisher 1992).

Summary and Closing Remarks

We have examined two examples of abrupt climatic transitions from the Paleogene, one involving global warming in the latest Paleocene, and the other cooling and glaciation in the earliest Oligocene. Both events occurred when global climates were in long term transition, both produced transient climate states, in one case a warm, nearly isothermal ocean, and the other, a full scale continental ice-sheet, and both were synchronous with important evolutionary changes in biota. Evidence for coeval changes in marine productivity and carbon cycling, and the lack of evidence for external forcing, support the idea that these transient climates resulted from feedbacks involving perturbations in the cycling of carbon between the oceans, atmosphere, and sediments. These feedbacks were most likely triggered by abrupt reorganizations of ocean and atmospheric circulation that occurred as critical thresholds were passed in the climate system.

The possibility that sudden reversals in ocean circulation might occur in response to gradual changes in global climate has been known for some time. This idea was impressed upon Chamberlin (1906) who in discussing deep ocean circulation noted *"that the battle between temperature and salinity is a close one, and that no profound change is necessary to turn the balance."* Although tremendous advances have been made toward a full understanding of ocean and climate dynamics in the nine decades since, the potential role of ocean feedbacks, physical and chemical, to modify climate transitions is just beginning to be understood. Although the scales may differ, the geologic record demonstrates that prehistoric climate transitions were often abrupt, and sometimes resulted in transitional climates. By better defining the nature of these events, important insight may be gained into the feedback processes which may amplify or dampen rapid climate transitions.

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