

Chapter 7

Global warming leads the carbon isotope excursion at the Paleocene-Eocene thermal maximum

The prominent negative carbon isotope excursion (CIE) at the Paleocene-Eocene thermal maximum (55 Ma) is generally accepted to reflect a transient, massive input of isotopically light carbon into the ocean-atmosphere system. Many authors have assumed that this carbon led to pronounced global greenhouse warming. Here we show, from an expanded record in New Jersey, that both the onset of the global abundance of the subtropical dinoflagellate *Apectodinium* and surface-ocean warming as recorded by TEX₈₆ preceded the CIE by several thousands of years. The offset between *Apectodinium* and the CIE was confirmed in other sites from New Jersey, the North Sea and New Zealand. The ~3 kyrs time lag between the onset of warming and the CIE is consistent with the expected lag between bottom water warming and submarine methane hydrate dissociation, suggesting that the latter mechanism indeed caused the CIE.

Introduction and material

The idea that climate change during the Paleocene-Eocene thermal maximum (PETM, at ~55.5 Ma) started simultaneously with the CIE comes from the many stable isotope records from deep-sea sediments, which often show an excursion in stable oxygen isotopic composition ($\delta^{18}\text{O}$) of biogenic carbonate concomitant with the CIE (Zachos et al., 2001). Some climate proxy records suggest that some warming and biotic changes slightly predated the CIE (Bowen et al., 2001; Thomas et al., 2002; Tripathi and Elderfield, 2005), but it has been argued that such evidence for pre-CIE warming is within the range of normal variability (Cramer and Kent, 2005). If pre-CIE warming did occur, it would suggest that some initial warming somehow triggered the injection of ^{13}C -depleted carbon (Dickens et al., 1995; Thomas et al., 2002). Unraveling the sequence of events on millennia-scale is difficult from deep marine sediment sections because they often suffer from severe carbonate dissolution (Chapter 1) and/or are too condensed and/or bioturbated across the PETM. Marine successions that would potentially circumvent these problems as a result of high sediment accumulation rates are located in the neritic regions. To unravel the relative and absolute timing of the various geochemical and biotic events associated with the PETM, we generated data at locations where sections represent neritic deposition.

For this purpose, we used Ocean Drilling Program Leg 174AX Site 'Bass River' and United States Geological Survey (USGS) borehole 'Wilson Lake' from the New Jersey Shelf at ~40°N paleolatitude, and the recently released industry well FINA 30 14-1 from the North Sea at ~55°N paleolatitude (Fig. 1). We carried out palynology, and where possible, we combined these data with TEX_{86} paleothermometry and $\delta^{18}\text{O}$ records of bulk carbonate and planktonic foraminifera (Chapter 4; John et al., in prep), in order to elucidate the pattern of environmental change in relation to the CIE across the PETM. Then we compared the results with previously published records from the ACEX cores from the Arctic Ocean at ~85°N paleolatitude and the Tawanui section (Chapter 3), New Zealand at ~55°S paleolatitude (Crouch et al., 2001). The CIE was previously identified in the New Jersey Shelf cores by a negative step in stable carbon isotopic composition ($\delta^{13}\text{C}$) of bulk carbonate ($\delta^{13}\text{C}_{\text{BC}}$) and foraminifera ($\delta^{13}\text{C}_{\text{TOR}}$) (Cramer et al., 1999; Gibbs et al., 2006; John et al., in prep; Chapter 4), and we amended these data by generating $\delta^{13}\text{C}$ records on organic walled dinoflagellate cysts (dinocysts; $\delta^{13}\text{C}_{\text{DINO}}$) (Figs. 2, 3). At the North Sea site, we identified the CIE based on $\delta^{13}\text{C}$ records of total organic carbon ($\delta^{13}\text{C}_{\text{TOC}}$) (Fig. 3) and by the presence of the dinoflagellate *Apectodinium augustum*, which is diagnostic of the PETM (Bujak and Brinkhuis, 1998; Crouch et al., 2001). In the Arctic Ocean (Chapter 3) and New Zealand (Kaiho et al., 1996; Crouch et al., 2001), the CIE and associated *Apectodinium* acme were previously identified. All our sites yield rich assemblages of palynomorphs, notably dinoflagellate cysts. In

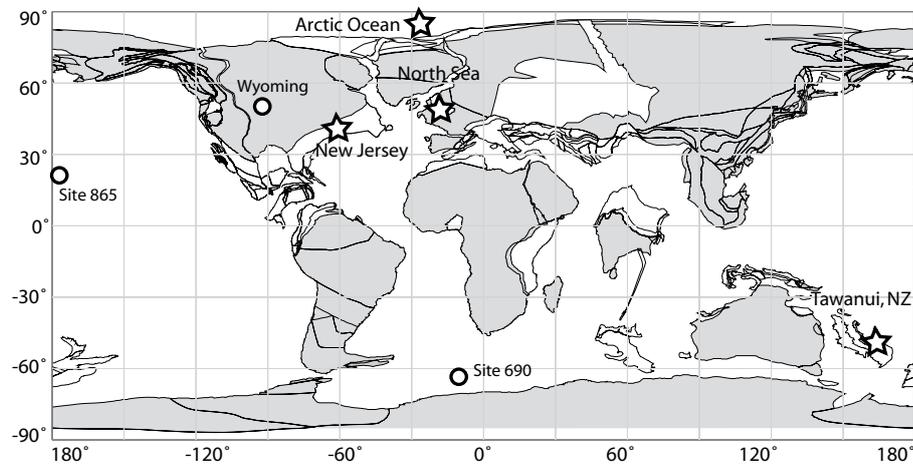


Figure 1. Location of the studied sites (stars) and sites with previously published records (circles; see text for references) within a paleogeographic reconstruction of the Earth at PETM times (source: <http://www.odsn.de/odsn/services/paleomap/paleomap.html>). Estimated regions of submerged continental shelves are indicated by black lines surrounding white areas.

addition, the New Jersey successions contain calcareous microfossils (Cramer et al., 1999; Gibbs et al., 2006; John et al., in prep; Chapter 4).

The stratigraphic thickness of the CIEs comprises 10.5 m at Bass River, 13.5 m at Wilson Lake, and approximately 37 m in the North Sea site (Appendix Figs. 1 and 2). Assuming a duration of 170 kyrs for the CIE (Röhl et al., in prep), estimates of average accumulation rates across the CIE are 6 cm.kyr⁻¹ for Bass River, 8 cm.kyr⁻¹ for Wilson Lake and 21 cm.kyr⁻¹ for the North Sea section. Both New Jersey sections yield potentially higher sedimentation rates because the upper boundary of the PETM represents a sea level driven unconformity at these sites (Cramer et al., 1999; Gibbs et al., 2006; Chapters 4, 6). Through the CIE at Bass River, potential precession related cycles can be recognized in the percentage *Apectodinium* of the dinocyst assemblage and number of dinocysts per gram of sediment, which would imply that sedimentation rates within the CIE are ~10 cm.kyr⁻¹ (Appendix 5). Assuming that and constant sedimentation rates through the studied intervals, this implies that temporal sample spacing of these records near the onset of the CIE comprises ~1 kyr for Bass River, ~2 kyr for Wilson Lake and ~3 kyr for the North Sea site, which is 5 to 20 times higher than reached at many deep sea sites (Bralower et al., 1997; Zachos et al., 2003; Tripathi and Elderfield, 2005); Chapter 1). The upper bathyal Tawanui section in New Zealand exhibits significantly lower sedimentation rates (Kaiho et al., 1996; Crouch et al., 2001), while the CIE in the Arctic Ocean record is located in a core gap (Chapter 3).

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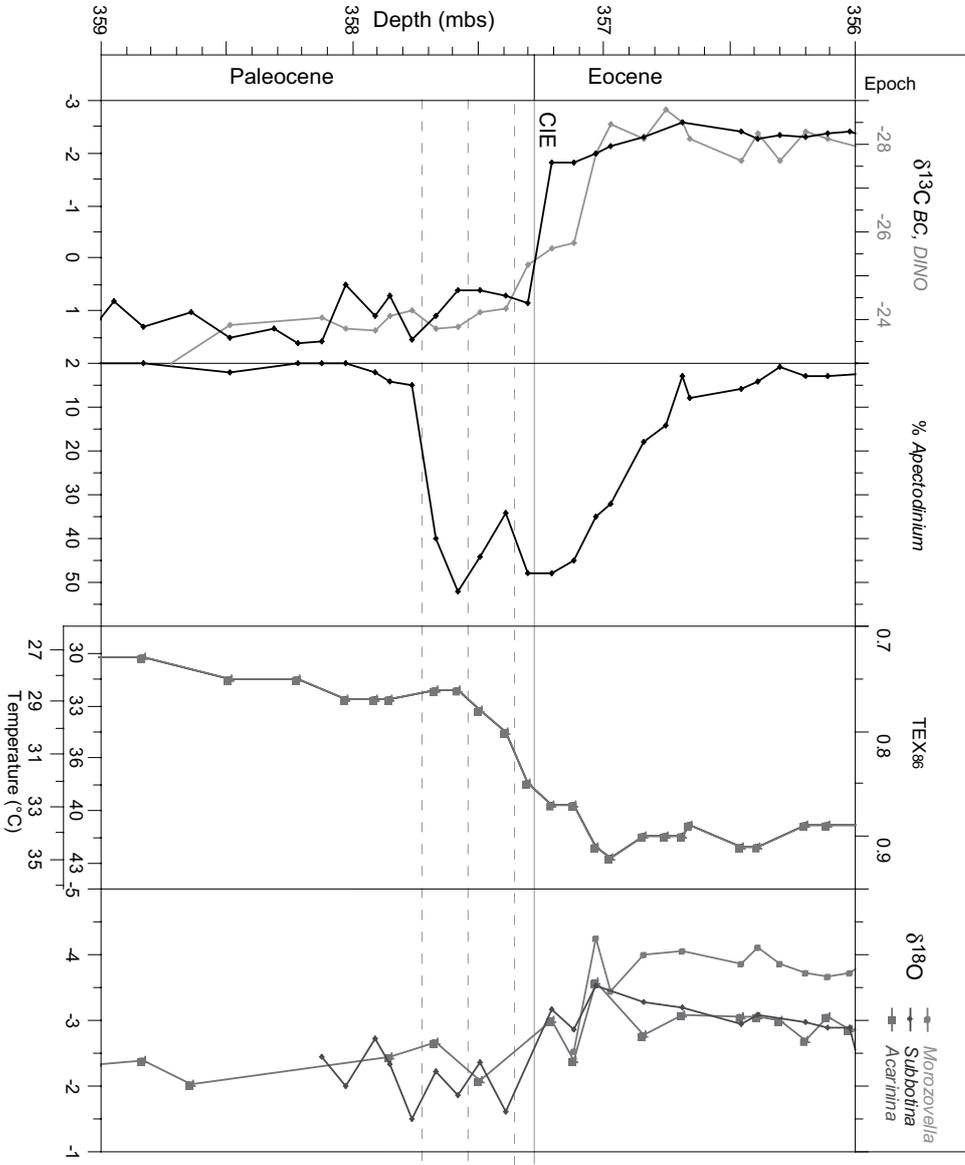


Figure 2. High-resolution records across the onset of the PETM at Bass River, New Jersey. BC = bulk carbonate, DINO = dinocysts, VPDB = Vienna Pee Dee Belimnite, mbs = meters below surface. Scales at TEX₈₆ temperatures represent calibrations by Schouten et al. (2002) for the top bar and by (Schouten et al. (2003) for the lower bar. Stable isotope data on carbonate are from John et al. (in prep).

Results

At Bass River, the sharp decrease in both $\delta^{13}\text{C}_{DINO}$ and $\delta^{13}\text{C}_{BC}$ occurred at 357.3 meters below surface (mbs) (Fig. 2). There is some indication of a decline in carbon isotope values below this increase, but this may be within the range of background variability. Hence, we identify the base of the CIE, which marks the Paleocene-Eocene boundary, at 357.3 mbs. The onset of the globally recorded (Crouch et al., 2001; Chapters 3 and 5) acme of the dinocyst *Apectodinium* (at 357.7 mbs) precedes the CIE by 40 cm. This cannot be due to bioturbation because the CIE is actually identified on dinocysts through an interval where assemblages are relatively constant. Assuming above sedimentation rates, the onset of the *Apectodinium* acme leads the onset of the CIE by in the order of 4-5 kyr.

To assess surface ocean temperatures at the New Jersey shelf, we used the multi-specimen planktonic foraminifer $\delta^{18}\text{O}$ records of (John et al., in prep) and applied the organic paleothermometer TEX_{86} at the Bass River site. The $\delta^{18}\text{O}$ records exhibit much scatter but do show lower values during the PETM than before. The negative excursions in the surface dweller *Acaranina* ($\sim -0.5\text{‰}$) and thermocline dweller *Subbotina* ($\sim -1\text{‰}$) are rather small. Due to the scatter it is hard to pinpoint the stratigraphic level at which the values start to decrease. Critically, not enough foraminifera were present in the sample at 357.30 mbs, potentially associated to dissolution, prohibiting $\delta^{18}\text{O}$ analysis. It could be argued that several planktonic foraminifer $\delta^{18}\text{O}$ values just below the onset of the CIE indicate warming. TEX_{86} is based on the relative abundance of crenarchaeotal membrane lipid isomers and is independent of surface water parameters such as nutrient availability or salinity. Moreover, it shows a highly significant linear correlation with present-day mean annual SST (Schouten et al., 2002; Wuchter et al., 2004). The TEX_{86} values measured at Bass River exceed the range of modern box-core calibrations. Therefore, the inferred absolute temperatures ($^{\circ}\text{C}$) should be treated with care. However, the magnitude of warming during the PETM in TEX_{86} at Wilson Lake is matched by the planktonic foraminifer $\delta^{18}\text{O}$ record (Chapter 4). Terrestrially-derived lipids are present in insignificant amounts ($\text{BIT} = <0.15$ in all samples, Chapter 6), so they did not influence our TEX_{86} records. The TEX_{86} record from Bass River evidences gradual warming, showing intermediate temperature values between pre- and mid-PETM, which are extremely rare in single-specimen foraminifera-derived $\delta^{18}\text{O}$ values. Critically, the record implies that the onset of anomalous warming is located $\sim 25\text{-}30$ cm below the CIE, with approximately half of the PETM warming occurring before the onset of the CIE (Fig. 2). Assuming above sedimentation rates, the onset of warming preceded the onset of the CIE by approximately 2.5-3 kyrs (Fig. 2).

At Wilson Lake, the onset of the CIE, representing the main negative step in $\delta^{13}\text{C}$, is assigned to 109.8 mbs based on both $\delta^{13}\text{C}_{DINO}$ and $\delta^{13}\text{C}_{BC}$ (Fig. 3). This implies that also here the onset of the *Apectodinium* acme (~ 110.4) leads the CIE

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by ~0.6 m. Assuming the above sedimentation rates this would represent about 4-5 kyr. Although less clear than at Bass River, the TEX_{86} record at Wilson Lake also suggests temperature rise prior to the CIE (fig. 4 in Chapter 4, p. 58). The foraminifer $\delta^{18}\text{O}$ records show a similar pattern, although with more scatter. At the North Sea site, the lowermost position to which the onset of the CIE can be attributed is at ~2927 meters below sea floor (mbsf) (Fig. 3), while the onset of the *Apectodinium* acme is at ~2927.5 mbsf. Hence, the onset of the acme leads the $\delta^{13}\text{C}_{\text{TOC}}$ CIE by ~0.5 m, which represents about 4 kyr at this site. Critically, this site is located in the Central North Sea basin, so sedimentation rates can be assumed more constant than at the shelf sections in New Jersey. At Tawanui, New Zealand, the base of the CIE is more difficult to locate; based on the slightly scattered $\delta^{13}\text{C}_{\text{TOC}}$ it could be placed at 3 cm (Fig. 3), but the $\delta^{13}\text{C}_{\text{BC}}$ record suggests the CIE starts at 4 cm. Either way, the onset of the *Apectodinium* acme is at 2 cm, indicating that also here the onset of the acme precedes the CIE. Due to the condensed nature of this record and the limited stratigraphic offset, estimation of the time lag is hampered. Together, our records indicate that conditions causing the global acme of this species were not directly related to the onset of the CIE. Significantly, our data show that dinoflagellate assemblages did not change much across the onset of the CIE globally; yet they rank among the ecologically most sensitive plankton (Appendix 1). This suggests that the input of ^{13}C -depleted carbon reflecting the CIE may not have caused large environmental perturbations. Interestingly, the TEX_{86} data from the Arctic Ocean PETM record appears to show anomalous warming in the uppermost Paleocene sample, hence before the CIE (fig. 2 in Chapter 3, p. 44), corroborating the observations at Bass River and perhaps Wilson Lake. The time lag involved at this site is unknown due to the core gap in between the uppermost Paleocene and the PETM (Chapter 3), which hampers confident correlation to the sites on the New Jersey Shelf. Abundant *Apectodinium* is not recorded in this sample, suggesting that this taxon only reached the Arctic Ocean after the initial warming.

Discussion

In our records, the onset of the *Apectodinium* acme is the earliest sign of anomalous environmental change associated with the PETM. Identification of the environmental parameters that caused the acme is therefore vital in understanding the sequence of climatological events that eventually caused the warming and the CIE. Crouch et al. (2003a), p. 125) note that any *Apectodinium* bloom required “a special set of environmental conditions” of which a baseline requirement appears to be high temperatures. *Apectodinium* acmes have been recorded from upper Paleocene deposits in the Tethyan Ocean, suggesting that conditions there were episodically and locally similar to those on a global scale during the PETM (Crouch et al., 2003a). Similar to other mid-latitude regions, *Apectodinium* was already present on the New Jersey shelf at least since Chron

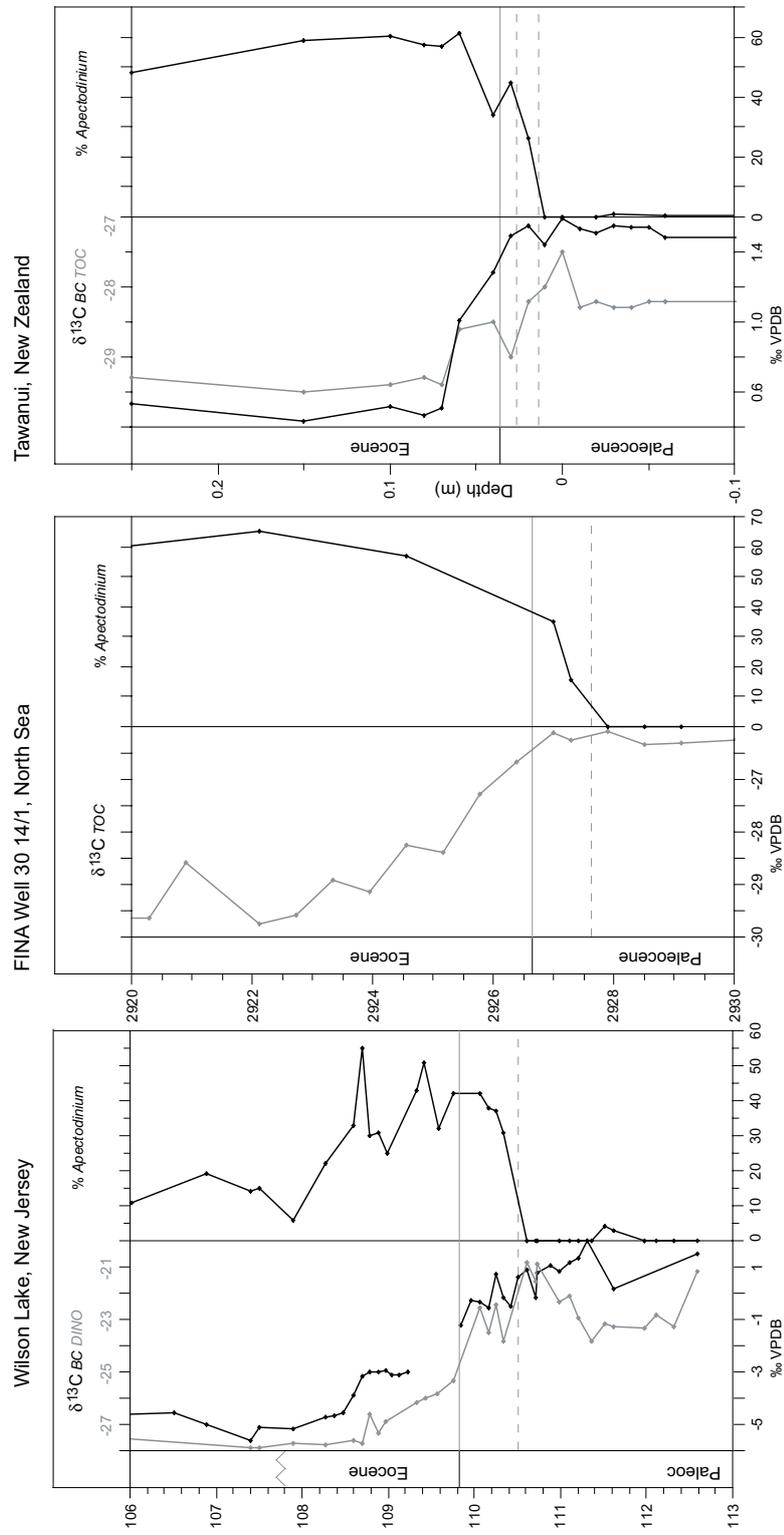


Figure 3. High-resolution records across the onset of the PETM at Wilson Lake, New Jersey (bulk carbonate stable isotope data from Chapter 4), FINA Well 30 14/1, North Sea and Tawanui, New Zealand (data from Crouch et al., 2001). BC = bulk carbonate, DINO = dinocysts, VPDB = Vienna Pee Dee Belimnite, mbs = meters below surface.

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C25n (Appendix Fig. 5.1); yet, in contrast to equatorial sites no pre-PETM acmes have been reported from such regions. Since *Apectodinium* was abundant in the Arctic Ocean with SSTs around 23°C (Chapter 3), New Jersey shelf SSTs during the late Paleocene should have already been high enough to allow for abundant *Apectodinium*. This implies that some other environmental parameter(s) prevented the establishment of late Paleocene *Apectodinium* acmes in the mid latitudes. It has also been noted that *Apectodinium* locally became outnumbered by typical low-salinity tolerant dinocysts during the PETM (Chapter 3). This observation is consistent with the records from the New Jersey shelf (Appendix 5), indicating that very low salinities were not optimal for *Apectodinium*. Other proposed ecological requirements include stratified surface waters (Crouch et al., 2003a). Moreover *Apectodinium* has morphological characteristics identical to cysts of modern heterotrophic dinoflagellates, which has fueled the hypothesis that *Apectodinium* was a heterotrophic dinoflagellate (Bujak and Brinkhuis, 1998). Basic predator-prey abundance models indicate that with higher nutrient supplies, ecosystems should become relatively enriched in organisms that are higher up in the food chain, e.g., heterotrophic. The total amount of dinoflagellate cysts per gram of sediment, which reflects cyst production and thereby nutrient supply during the PETM at Bass River, covaries absolute abundance of *Apectodinium* cysts (Appendix Fig 5.3). This suggests that higher nutrient levels are directly reflected in higher production of *Apectodinium* cysts, supporting the hypothesis that *Apectodinium* was a heterotrophic dinoflagellate. Increasing nutrient levels may, therefore, have contributed to the *Apectodinium* acme. If so, the global character of the acme implies that at least neritic sections underwent significant eutrophication on a global scale, a hypothesis corroborated by many proxy data (Chapter 5). Modern dinoflagellate blooms usually last for several days to weeks (Dale, 1996). Conceivably, *Apectodinium* blooms during the PETM had similar dynamics, in which case the pre-CIE signal would imply a change in specific seasonal conditions of the surface waters. This may include any of the above environmental factors. However, even a combination of these factors was likely not truly unique in the early Paleogene, suggesting that some critical environmental factor has not yet been identified. Whichever combination of surface water parameters caused the global acme of *Apectodinium*, it is certainly associated with the PETM and appears to signify an ‘early warning’ to global warming.

Our data convincingly show that the onset of the *Apectodinium* acme and the beginning of the anomalous warming, both intrinsically related to the PETM, are themselves not directly related to the input of ¹³C-depleted carbon that caused the CIE. This observation provides a new perspective to the global change that was associated with the PETM. Potentially, the pre-CIE warming was caused by a large increase of a non-carbon greenhouse gas, which would not leave a signature on our $\delta^{13}\text{C}$ records. Alternatively, if the pre-CIE warming was forced by an increase in atmospheric carbon, its magnitude of 2-3°C requires approximately

a doubling of atmospheric CO₂, depending on uncertainties regarding the climate sensitivity (Archer, in press). If so, the lack of a signal in the δ¹³C curves implies that this carbon source had the same isotopic composition as the late Paleocene exogenic carbon pool, suggesting that the ocean may have been the source of atmospheric carbon increase. Mantle carbon has been suggested to have caused initial warming (Dickens et al., 1995; Bralower et al., 1997; Schmitz et al., 2004; Thomas and Bralower, 2005), but given its δ¹³C range between -5 to -7‰, this should have led to a slight negative excursion in the δ¹³C records.

A relatively popular hypothesis to explain the CIE is the injection of ¹³C-depleted carbon through the dissociation of large amounts of submarine methane clathrates (Dickens et al., 1995). In the present ocean, these hydrates are stable along continental slopes at relatively high pressure and low temperatures (Kvenvolden, 1988). It has been argued that during the much warmer latest Paleocene the reservoir was smaller, because hydrates were not stable at the relatively shallow waterdepths where much clathrates are present in the modern ocean (Cramer and Kent, 2005; Higgins and Schrag, 2006). However, methane hydrates were potentially present at greater depths than at present (Dickens, 2001c). This would have required a larger supply of organic matter further away from the continents than nowadays. In conjunction, less organic matter was subject to oxic degradation due to the lower oxygen content of the bottom waters due to the higher temperatures, potentially enhancing anaerobic degradation and methane clathrate production. This suggests that a sufficient reservoir of methane clathrates to generate the CIE was potentially present during the latest Paleocene. If so, interestingly, the time scale for thermal destabilization of methane hydrates is in the order of thousands of years (MacDonald, 1990; Xu et al., 2001), which is exactly in accordance with the time lag we record between warming and the CIE.