

Rapid carbon injection and transient global warming during the Paleocene-Eocene thermal maximum

A. Sluijs* & H. Brinkhuis

Palaeoecology, Institute of Environmental Biology, Utrecht University, Laboratory of Palaeobotany and Palynology, Utrecht, the Netherlands.

* Corresponding author. Email: A.Sluijs@uu.nl

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Introduction

The Paleocene–Eocene Thermal Maximum (PETM), ~55.5 Myr ago, was a geologically brief (~170 kyr) episode of globally elevated temperatures, which occurred superimposed on the long-term late Paleocene and early Eocene warming trend (Fig. 1). It was marked by a 5–8°C warming in both low and high-latitude regions, a perturbation of the hydrological cycle and major biotic response on land and in the oceans, including radiations, extinctions and migrations (see overviews in Bowen et al., 2006; Sluijs et al., 2007a). In addition, the PETM is associated with a pronounced negative carbon isotope excursion (CIE), recorded as a >2.5‰ decrease in the stable carbon isotope composition ($\delta^{13}\text{C}$) of sedimentary components (e.g., Kennett and Stott, 1991; Koch et al., 1992) (Fig. 1). Although there is discussion on the exact magnitude of the CIE, as it appears to depend on substrate (e.g., Schouten et al., 2007), it can only be explained by a massive (at least 1.5×10^{18} g; 1500 Gt) injection of ^{13}C -depleted carbon into the ocean-atmosphere system (Dickens et al., 1995; 1997).

Recent work has focused on elucidating the injection mechanism(s) and volume of the carbon that caused the CIE, but has also addressed the question whether the ^{13}C -depleted carbon caused the warming or acted as a positive feedback in an already warming world. Moreover, was the PETM a unique event in the early Paleogene greenhouse world and what is the relevance of the PETM for current carbon injection into the atmosphere and/or the oceans from fossil fuel burning?

Ocean acidification

Analogous to the modern situation, the injection of a large mass of CO_2 or CH_4 (which would have been oxidized to CO_2 within a century at most; Schmidt and Shindell, 2003) should

have increased the acidity of the ocean. As a result, a shallowing of the calcite compensation depth (CCD) and dissolution of deep-sea carbonates should have occurred, thereby buffering the seawater pH change (Dickens et al., 1997). Indeed, the dissolution of deep sea carbonates has been documented in various deep ocean basins, based on the occurrence of clay layers as well as biogenic calcite fragmentation (e.g., Zachos et al., 2005) (Fig. 1). The severity of dissolution, however, appears to have been highly variable between various basins, perhaps caused by spatial variability in bioturbation (Zeebe & Zachos, 2007; Panchuk et al., 2008). Moreover, the magnitude of dissolution should have been equivalent with the amount of injected carbon and has, as such, been used to elucidate the source and volume of injected carbon by modeling (Panchuk et al., 2008). According to that study nearly 7000 Gt of carbon, derived from multiple sources, was injected into the ocean/atmosphere system during the PETM.

Carbon sequestration

The distribution of deep-sea carbonate abundances also points to one mechanism of carbon sequestration. Carbonate accumulation rates at many sites appear to have been very high towards the termination of the PETM – the lysocline (the depth in the ocean below which the rate of dissolution of calcite increases dramatically) was located even deeper than prior to the PETM – resulting in sequestration of large amounts of carbon (e.g., Kelly et al., 2005; Zachos et al., 2005). This phenomenon is model-predicted (Dickens et al., 1997) and was probably driven by silicate weathering (e.g., Ravizza et al., 2001), which slowly recharged the ocean with carbonate ion and eventually led to carbonate ion over-saturation and extremely good preservation of calcite on the sea floor. Recent work has revealed a potentially critical contribution of carbon

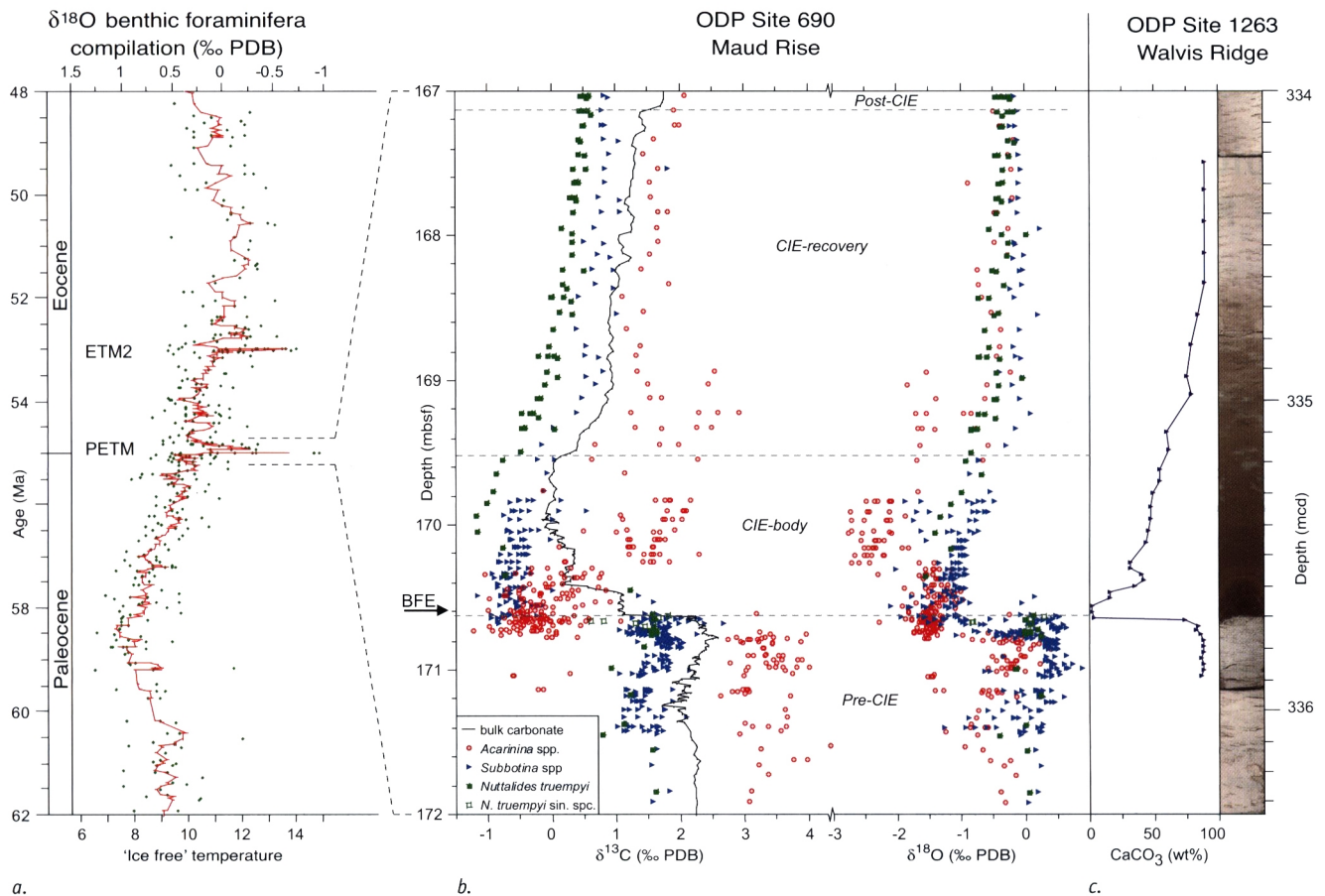


Fig. 1. a. Benthic foraminifer $\delta^{18}\text{O}$ record from 62 - 48 Myr (adapted from Zachos et al., 2001; data for ETM2 represent the *Cibicoides* record from Lourens et al., 2005). $\delta^{18}\text{O}$ values represent corrected values (cf. Zachos et al., 2001). An additional correction of -0.25‰ was applied to the ETM2 *Cibicoides* values to synchronize baseline levels between the records. ETM2 = Eocene Thermal Maximum 2 (also referred to as H1 or Elmo), PETM = Paleocene-Eocene Thermal Maximum; b. Compilation of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values across the PETM of planktonic foraminifera (surface dweller *Acarinina* and thermocline dweller *Subbotina* spp.; mostly single specimen), benthic foraminifera (*Nuttallides truempyi*) and bulk carbonate from ODP Site 690 from Maud Rise in the Weddell Sea. Data are from Kennett & Stott, 1991; Bains et al., 1999; Thomas et al., 2002, Kelly et al., 2005 and the figure was modified from Sluijs et al., 2007a); c. Carbonate weight % record and core photo of ODP Site 1263 at Walvis Ridge, southeastern Atlantic (adapted from Zachos et al., 2005), showing dissolution and resulting clay layer. BFE refers to the main phase of benthic foraminifer extinction, mbsf = meters below sea floor; mcd = meters composite depth.

sequestration on continental shelves (Sluijs et al., 2008; John et al., 2008). At several sites, calcite burial rates increased (John et al., 2008). Moreover, particularly in the Arctic but also in the Tethys, organic carbon burial increased during the PETM due to increased river runoff causing more organic production, as well as stratification and bottom water anoxia (Sluijs et al., 2008). Excess burial perhaps comprised in the order of 800 Gt of carbon during the PETM in the Arctic alone (Sluijs et al., 2008).

Meridional temperature gradients

The application of the organic paleothermometer TEX_{86} , as well as oxygen isotope data (Fig. 1) on well-preserved foraminifera, has recently led to a much better quantification of sea surface temperatures (SSTs) across the PETM. Background late Paleocene

and early Eocene temperatures were already warm, with SSTs of $\sim 32^\circ\text{C}$ in the tropics (Pearson et al., 2007), although the recently revised calibration of the TEX_{86} proxy, suggests that tropical temperatures were several degrees higher. Mid- and high-latitude surface ocean were approximately 25°C and 17°C , respectively (e.g., Sluijs et al., 2006, 2007b; Zachos et al., 2006). During the PETM, tropical as well as mid-latitude and Arctic SSTs rose by $5 - 8^\circ\text{C}$ (Zachos et al., 2003; Sluijs et al., 2006; Zachos et al., 2006) (Fig. 2). Such temperatures in the high Arctic are supported by biogeographical data, such as the abundant occurrence of subtropical dinoflagellates (Sluijs et al., 2006), and high values of a biomarker-based proxy for continental air temperatures (Weijers et al., 2007). Hence, meridional temperature gradients were significantly smaller during both background and PETM conditions, although it remains unclear if the Arctic data represent mean annual or

summer temperatures (Sluijs et al., 2006). Yet even if they represent summer temperatures and tropical temperatures even higher than previously estimated (Pearson et al., 2007), current generation fully-coupled climate models overestimate meridional gradients, even when the model is fed with Eocene geography and high CO₂ concentrations (Huber and Nof, 2006) (Fig. 2). This mismatch suggests that higher-than-modern greenhouse gas concentrations must have operated in conjunction with feedback mechanisms that either amplified polar temperatures or cooled the tropics and that are not incorporated in the models (Sluijs et al., 2006). Potential mechanisms include polar stratospheric clouds (Sloan and Pollard, 1998) and hurricane-induced ocean mixing (Emanuel et al., 2004; Sriver and Huber, 2007), for polar warming and tropical cooling, respectively.

Interestingly, the meridional temperature gradient did not further decrease during the PETM. This can be partly explained by the absence of ice-albedo feedbacks, since the Arctic was already ice-free prior to the PETM. Additionally, it implies that the mechanism that caused the reduced meridional temperature gradient did not become amplified during the PETM (Sluijs et al., 2006).

Additional early Eocene Hyperthermals

Recent work shows that a similar phase of extreme warming occurred at ~53.5 Myr (Lourens et al., 2005) (referred to in the literature as H-1, *Elmo* or Eocene Thermal Maximum 2; ETM2), and possible additional related phases at ~53.1 Myr (called I-1) and ~52.3 Myr (called K or X) (Cramer et al., 2003; Röhl et al., 2005; Nicolo et al., 2007). Although documentation of these phases is, as yet, relatively incomplete, the available

information indicates that these additional hyperthermals are also associated with massive injection of ¹³C-depleted carbon, ocean acidification and perturbations of the hydrological cycle, though less pronounced than during the PETM. Orbital tuning of the complete late Paleocene and early Eocene record at Walvis Ridge (South Atlantic) has indicated a link between the timing of the hyperthermals and eccentricity maxima (Lourens et al., 2005; Westerhold et al., 2007), which would have implications for the mechanisms that caused global change during the hyperthermals.

Leads and lags and mechanisms of carbon input

One prominent example of biotic change associated with the onset of the CIE is recorded along continental margins, where sediment sequences from all latitudes contain high abundances of dinoflagellate cysts belonging to the subtropical genus *Apectodinium* (Fig. 2) (Crouch et al., 2001; Sluijs et al., 2007a). In part, this must be associated to the PETM warming. However, in stratigraphically expanded marginal marine sections from the New Jersey Shelf and the North Sea, as well as a section in New Zealand, the onset of the *Apectodinium* acme started some 5 kyr prior to the CIE (Sluijs et al., 2007b) (Fig. 3). Additionally, the onset of the PETM SST warming at New Jersey appears to have led the CIE by several thousands of years (but lagged the onset of the *Apectodinium* acme) (Sluijs et al., 2007b). This disparity indicates that warm SST was not the only environmental control on *Apectodinium* abundances. Moreover, it suggests that the carbon release that caused the CIE was a result of initial climate change and acted as a positive feedback. This scenario fits the model of CH₄ release from submarine hydrates causing the CIE (Dickens et al., 1995). If this pre-CIE warming



Fig. 2. Two specimens of the dinocyst *Apectodinium augustum* from the PETM of Lomonosov Ridge, recovered during IODP Expedition 302.

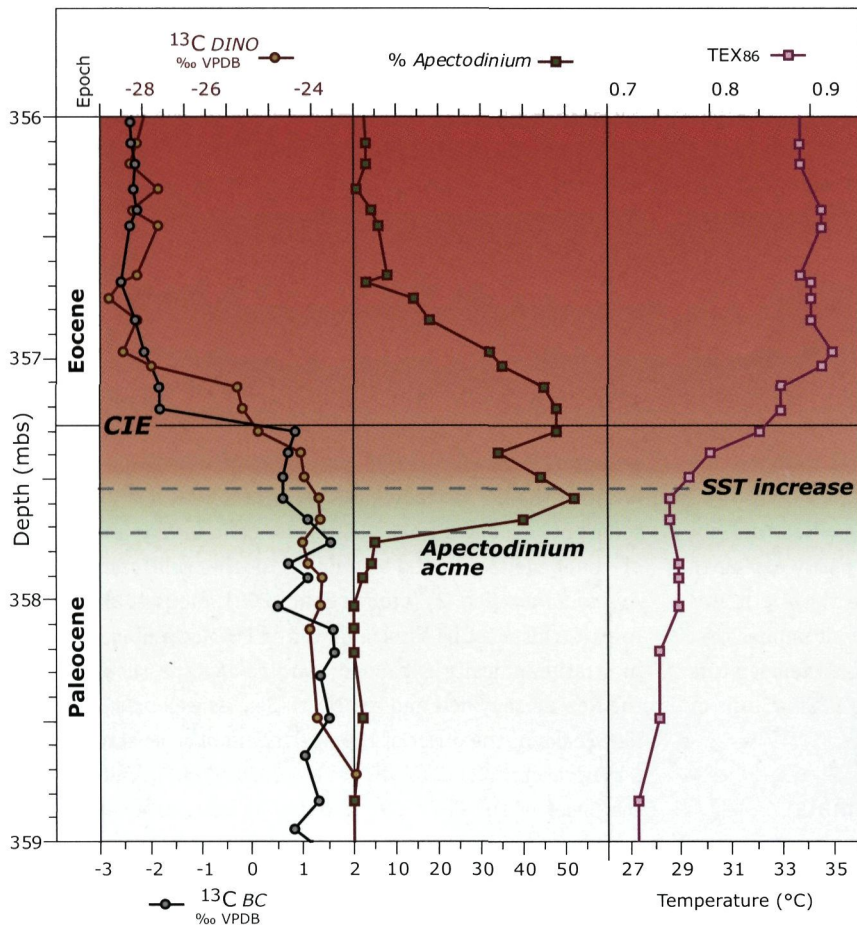


Fig. 3. High-resolution records across the onset of the PETM at Bass River, New Jersey Shelf Sites; redrawn from Sluijs et al., 2007b. Solid horizontal line at ~357.3 mbs represent the onset of the CIE; dashed lines represent the onsets of the Apectodinium acme and surface ocean warming. BC = bulk carbonate, DINO = dinocysts, VPDB = Vienna Pee Dee Belemnite, mbs = meters below surface

was global, it was likely induced by greenhouse forcing, suggesting that the PETM warming and ocean acidification were caused by at least two sources of carbon (Sluijs et al., 2007b).

Concluding remarks

The past years of research on the PETM and the newly discovered additional hyperthermals have resulted in a clearer picture of these critical phases in Earth's history. Improved drilling techniques (Integrated Ocean Drilling Program) have resulted in the recovery of complete sections, and new analytical techniques have contributed to much better quantitative estimates of surface temperatures. Moreover, high-resolution studies on expanded marginal marine sequences have identified leads and lags in the interaction between the climate system and the carbon cycle during the onset of the PETM. The new data provide fundamental constraints for modeling global climate and carbon cycling and are increasingly leading to a much better description and understanding of the state and dynamics of a (in this case the early Paleogene) greenhouse world.

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