SERIES

Geoscience of Climate and Energy 4.
Rapid Carbon Injection and Transient Global Warming during the Paleocene–Eocene Thermal Maximum

Appy Sluijs
Institute of Environmental Biology
Laboratory of Palaeobotany and Palynology
Utrecht University
Utrecht, The Netherlands
E-mail: a.sluijs@uu.nl

INTRODUCTION

The Paleocene–Eocene Thermal Maximum (PETM), ~55.5 My ago, was a geologically brief (~170 ky) episode of globally elevated temperature that was 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 and 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 ($\delta^{13}C$) of sedimentary components (e.g. Kennett and Stott 1991; Koch et al. 1992; Fig. 1). The CIE can only be explained by a carbon 'burp' – a massive (at least 1.5×10^{18} g or 1500 Gt) injection of ^{13}C-depleted carbon into the ocean-atmosphere system (Dickens et al. 1995).

Recent work has focused on elucidating the injection mechanism(s) and the quantity of carbon that caused the CIE, but has also addressed the question of whether the ^{13}C-depleted carbon caused the warming or acted as a positive feedback in an already warming world. Was the PETM a unique event in the early Paleogene greenhouse world, and what is the relevance of the PETM for current carbon injection 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 during the PETM has been documented in various deep-ocean basins, based on the occurrence of clay layers (the residual sedimentary component in the deep sea after calcite dissolution) 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 different basins, perhaps caused by spatial variability in bioturbation (Zeebe and Zachos 2007; Panchuk et al. 2008). Moreover, the magnitude of dissolution should have been equivalent to the amount of injected carbon and has, as such, been used in modelling experiments to elucidate the source and volume of injected carbon (Panchuk et al. 2008). According to that study, nearly 7000 Gt of carbon, derived from multiple sources, was injected during the PETM.

CARBON SEQUESTRATION

The distribution of deep-sea carbonate points to one mechanism of carbon sequestration during recovery of the PETM. Carbonate accumulation rates at many sites appear to have been very high in the latter stages 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 before the PETM, resulting in the 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, which slowly recharged the ocean with carbonate ions and eventually led to carbonate ion over-saturation and extremely good preservation of calcite on the sea floor. Another mechanism of carbon sequestration leading to the termination of the PETM comes from recent work that revealed a major increase in carbon burial rates on continental shelves during the event (Sluijs et al. 2008; John et al. 2008). At several sites, calcite burial rates increased (John et al. 2008), but increased organic carbon burial may have been significant as well. Particularly in the Arctic but also in the Tethys, organic carbon burial increased during the PETM due to increased river runoff, which caused increased organic production in the oceans, as well as stratification of the
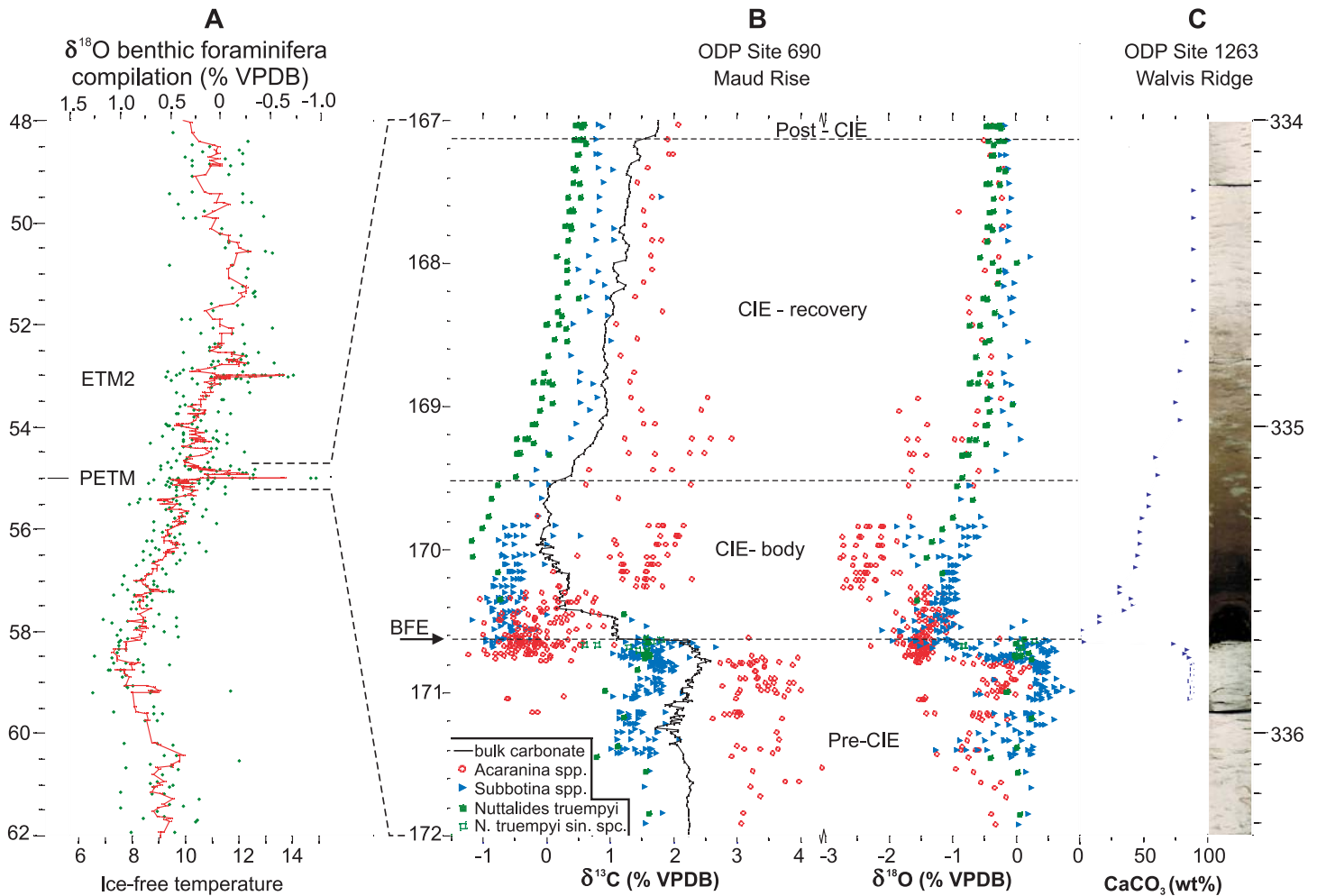
Figure 1. A) Benthic foraminifer δ18O record from 62-48 My, adapted from Zachos et al. (2001); data for ETM2 represent the *Cibicidoides* record from Lourens et al. (2005). The δ18O values (green diamonds) represent corrected values (cf. Zachos et al. 2001). The red line is a 5 point moving average. An additional correction of -0.25 ‰ was applied to the ETM2 *Cibicidoides* 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; VPDB: Vienna Pee Dee Belemnite δ18O and δ13C standard. B) Compilation of δ13C and δ18O 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 at Maud Rise in the Weddell Sea (data compilation and figure modified from Sluijs et al. 2007a). BFE refers to the main phase of benthic foraminifer extinction. CIE: carbon-isotope excursion. 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. mbsf: metres below seafloor; mcd: metres composite depth.

water column and bottom water anoxia (Sluijs et al. 2008). Excess burial perhaps amounted to roughly 800 Gt of carbon during the PETM in the Arctic Ocean alone (Sluijs et al. 2008).

MERIDIONAL TEMPERATURE GRADIENTS
The application of the organic paleothermometer TEX86 (based on the distribution of species of membrane lipids of the marine picoplankton *Crenarchaeota;* Schouten et al. 2002), as well as evidence from oxygen isotopes (Fig. 1) on well-preserved foraminifera, has

recently led to a much better quantification of sea surface temperatures (SSTs) across the PETM. Prior to the PETM excursion, 'background' Late Paleocene–Early Eocene was already warm, with SSTs of ~32 °C in the tropics (Pearson et al. 2007), although the recently revised calibration of the TEX86 proxy (Kim et al. 2008) suggests that tropical temperatures were several degrees higher. Mid- and high-latitude surface ocean temperatures 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°C (Zachos et al. 2003, 2006; Sluijs 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 other biogeochemical information (e.g. Weijers et al. 2007). Hence, meridional temperature gradients were significantly smaller than at present 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 higher than previously estimated (Pearson et al. 2007; Huber 2008), current-generation fully-coupled climate models overestimate Paleocene–Eocene meridional gradients, despite the model being fed with Eocene geography and high CO_2 concentrations (Huber and Nof 2006; Fig. 2). This 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, but that have not been incorporated in the models (Sluijs et al. 2006). Potential feedback mechanisms include polar stratospheric clouds (Sloan and Pollard 1998) and hurricane-induced ocean mixing (Emanuel et al. 2004; Sriver and Huber 2007), to account for polar warming and tropical cooling, respectively.

Interestingly, it appears that 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 warming phase occurred at ~53.5 My (referred to in the literature as H-1, Elmo or Eocene Thermal Maximum 2 (ETM2); Lourens et al. 2005), and possible additional phases at ~53.1 My (I-1) and ~52.3 My (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 hyperthermals are also associated with massive injection of ^{13}C-depleted carbon, ocean acidification and perturbations of the hydrological cycle, although 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.
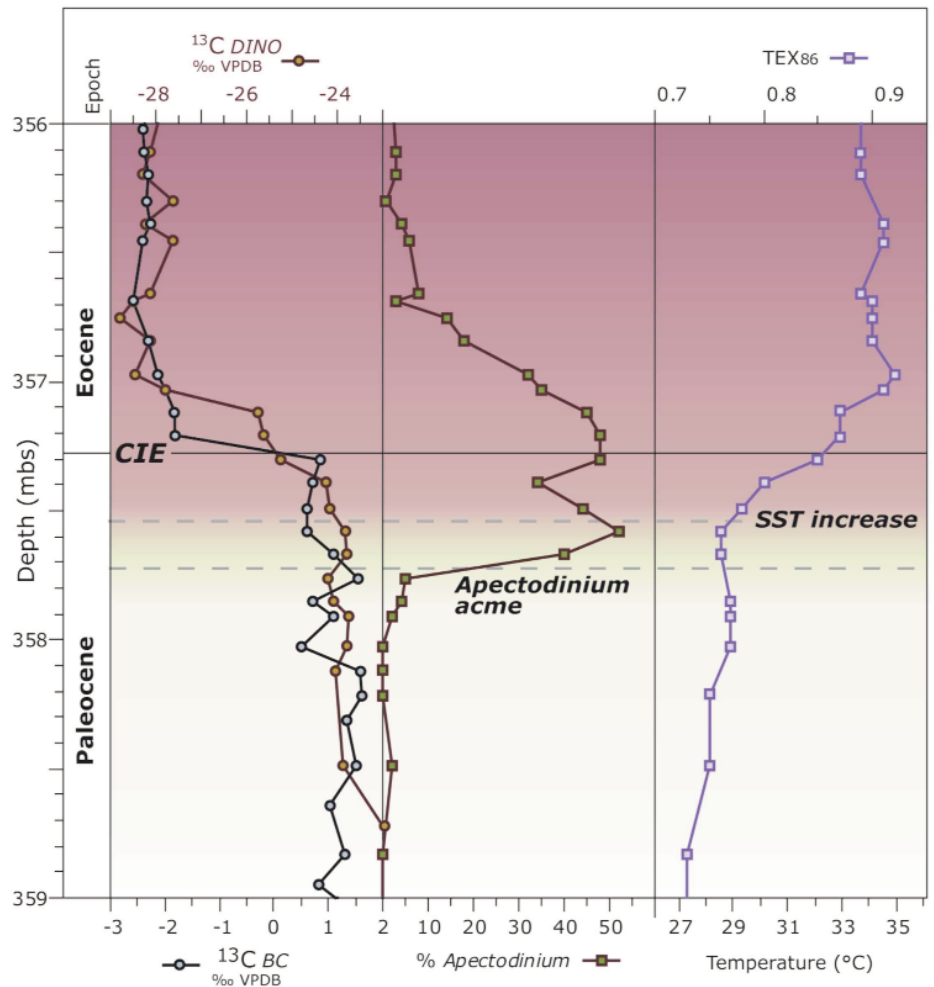

Figure 2. High-resolution records across the onset of the PETM at Bass River, New Jersey Shelf Sites; redrawn from Sluijs et al. (2007b). Solid horizontal line at ~357.3 mbs represents the onset of the carbon-isotope excursion (CIE); dashed lines represent the onsets of the *Apectodinium* acme and surface warming. BC: bulk carbonate; DINO: dinocysts; VPDB: Vienna Pee Dee Belemnite $\delta^{13}C$ standard; SST: sea-surface temperature; TEX$_{86}$: *Crenarcheota* (marine picoplankton) paleothermometer; mbs: metres below surface.

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 PETM is recorded along continental margins, where sedimentary sequences across the event from all latitudes contain high abundances of dinoflagellate cysts belonging to the subtropical genus *Apectodinium* (Crouch et al. 2001; Sluijs et al. 2007a). In part, this must be associated with 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 ky prior to the CIE (Sluijs et al. 2007b; Fig. 2). 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 indicates that warm SST was not the only environmental control on *Apectodinium* abundances. Moreover, it suggests that the carbon 'burp' that caused the CIE was a result of initial climate change and acted as a positive feedback. This scenario fits a model that relates the CIE to CH_4 release from submarine hydrates (Dickens et al. 1995). If this

pre-CIE warming 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 modelling 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, early Paleogene) greenhouse world.

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