

# Integrated stratigraphy of the Paleocene-Eocene thermal maximum in the New Jersey Coastal Plain: Toward understanding the effects of global warming in a shelf environment

Peter Stassen,<sup>1</sup> Ellen Thomas,<sup>2,3</sup> and Robert P. Speijer<sup>1</sup>

Received 19 March 2012; revised 18 October 2012; accepted 24 October 2012; published 22 December 2012.

[1] In the New Jersey Coastal Plain, a silty to clayey sedimentary unit (the Marlboro Formation) represents deposition during the Paleocene-Eocene thermal maximum (PETM). This interval is remarkably different from the glauconitic sands and silts of the underlying Paleocene Vincentown and overlying Eocene Manasquan Formation. We integrate new and published stable isotope, biostratigraphic, lithostratigraphic and ecostratigraphic records, constructing a detailed time frame for the PETM along a depth gradient at core sites Clayton, Wilson Lake, Ancora and Bass River (updip to downdip). The onset of the PETM, marked by the base of the carbon isotope excursion (CIE), is within the gradual transition from glauconitic silty sands to silty clay, and represented fully at the updip sites (Wilson Lake and Clayton). The CIE “core” interval is expanded at the updip sites, but truncated. The CIE “core” is complete at the Bass River and Ancora sites, where the early part of the recovery is present (most complete at Ancora). The extent to which the PETM is expressed in the sediments is highly variable between sites, with a significant unconformity at the base of the overlying lower Eocene sediments. Our regional correlation framework provides an improved age model, allowing better understanding of the progression of environmental changes during the PETM. High-resolution benthic foraminiferal data document the change from a sediment-starved shelf setting to a tropical, river-dominated mud-belt system during the PETM, probably due to intensification of the hydrologic cycle. The excellent preservation of foraminifera during the PETM and the lack of severe benthic extinction suggest there was no extreme ocean acidification in shelf settings.

**Citation:** Stassen, P., E. Thomas, and R. P. Speijer (2012), Integrated stratigraphy of the Paleocene-Eocene thermal maximum in the New Jersey Coastal Plain: Toward understanding the effects of global warming in a shelf environment, *Paleoceanography*, 27, PA4210, doi:10.1029/2012PA002323.

## 1. Introduction

[2] The early Paleogene was a climatically dynamic period, during which relatively short intervals of rapid global warming (hyperthermals) were superimposed on a warm background climate [Shuijs *et al.*, 2007a; Zachos *et al.*, 2008]. The Paleocene-Eocene thermal maximum (PETM), directly after the Paleocene-Eocene boundary, is the most extreme and best studied hyperthermal, with reconstructions of

global temperature anomalies available from the deep-sea to terrestrial environments [e.g., *McInerney and Wing*, 2011]. The ~170-kyr-long PETM [Röhl *et al.*, 2007] is linked to a perturbation of the global carbon cycle, as seen in a several per mille negative carbon isotope excursion (CIE) recorded globally, in marine and terrestrial settings [e.g., *Kennett and Stott*, 1991; *Koch et al.*, 1992; *Thomas and Shackleton*, 1996]. The CIE is caused by the rapid release of a large amount of <sup>13</sup>C-depleted carbon compounds into the atmosphere-ocean system, causing a rise in concentration of greenhouse gases and global warming. The approximate volume of released carbon compounds is still under debate, but has been commonly estimated between 2000 and 4500 gigaton [Dickens *et al.*, 1997; Zachos *et al.*, 2005; Dickens, 2011]. The rate of emission is also debated [Shuijs *et al.*, 2012], ranging from a few thousand years [Zachos *et al.*, 2007] to about 20 kyr [Cui *et al.*, 2011]. There is no agreement on source and method of emission of carbon compounds, hypotheses ranging from dissociation of submarine methane hydrates [e.g., Dickens *et al.*, 1997; Dickens, 2011] to burning of peat deposits [Kurtz *et al.*, 2003], to oxidation of marine organic matter due to igneous intrusion [Svensen *et al.*, 2004], or oxidation of

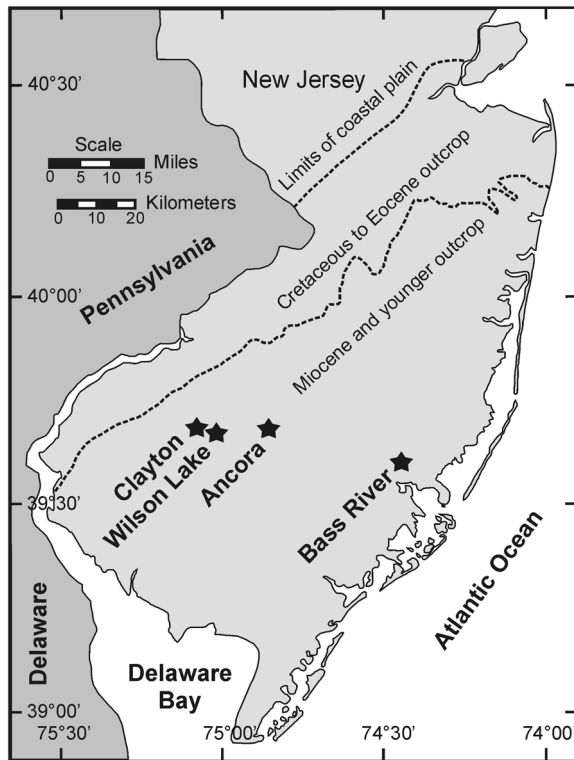
<sup>1</sup>Biogeology Research Group, Department of Earth and Environmental Sciences, KU Leuven, Leuven, Belgium.

<sup>2</sup>Department of Geology and Geophysics, Yale University, New Haven, Connecticut, USA.

<sup>3</sup>Department of Earth and Environmental Science, Wesleyan University, Middletown, Connecticut, USA.

Corresponding author: P. Stassen, Biogeology Research Group, Department of Earth and Environmental Sciences, KU Leuven, Celestijnenlaan 200E, BE-3001 Leuven, Belgium. (peter.stassen@ees.kuleuven.be)

©2012. American Geophysical Union. All Rights Reserved. 0883-8305/12/2012PA002323



**Figure 1.** Map with core sites [modified after Miller, 1997].

terrestrial organic matter in thawing permafrost [DeConto *et al.*, 2012], or some combination of carbon from different sources [Pagani *et al.*, 2006].

[3] Despite these uncertainties, the onset of the CIE is an outstanding global correlation level and has been used to define the base of the Eocene [Dupuis *et al.*, 2003; Aubry *et al.*, 2007]. The shape of the CIE is influenced by sedimentation rates, but provides tie points, allowing subdivision of the PETM in a CIE “core” (sustained low  $\delta^{13}\text{C}$  values) and recovery interval, with  $\delta^{13}\text{C}$  values increasing relatively rapidly during recovery phase I and more gradually in phase II [e.g., Röhl *et al.*, 2007]. The PETM triggered prominent changes in marine ecosystems [e.g., Speijer *et al.*, 2012]. Many biotic events can be used for worldwide correlation, specifically the extinction of deep-sea benthic foraminifera [Thomas, 1998] and the lowermost occurrence of distinct planktic foraminiferal [Kelly *et al.*, 1996; Guasti and Speijer, 2007] and nannoplankton (*Discoaster araneus* and *D. anartios*) associations, generally called ‘excursion taxa’, as well as poleward expansion of lower latitude taxa [Raffi *et al.*, 2005, 2009; Aubry *et al.*, 2007]. The potentially heterotrophic dinoflagellate cyst *Apectodinium* had an acme during the PETM associated with an expansion to higher latitudes, possibly influenced by changes in trophic state [Crouch *et al.*, 2001; Sluijs *et al.*, 2007a].

[4] An expanded PETM record in midlatitude shelf settings occurs in outcrops and boreholes along the northeastern coastal plains of the United States [Olsson and Wise, 1987; Gibson *et al.*, 1993; Gibson and Bybell, 1994], and has been studied extensively (see below). It is, however, not yet clear

how much of the time span of the PETM is represented in the sedimentary record. This question is addressed by integration of biostratigraphic, lithostratigraphic, ecostratigraphic and stable isotope records. We thus provide a new detailed regional correlation framework, with significant differences from published records [e.g., Gibbs *et al.*, 2006a] and is the basis for understanding of the environmental and biotic responses to the PETM.

## 2. Geologic Setting and Previous Studies

[5] The New Jersey Coastal Plain (NJCP) in the northern part of the Salisbury Embayment contains one of the most complete North Atlantic records of shallow marine Paleocene-Eocene boundary deposition in America [Olsson and Wise, 1987; Gibson *et al.*, 1993; Gibson and Bybell, 1994]. Sediments dominantly consist of bioturbated glauconitic clays, silts and sands, deposited on a sediment-starved siliciclastic shelf [Browning *et al.*, 2008]. The upper Paleocene Vincentown Formation (Fm.) consists of green-to-brown, glauconitic quartz sands updip (closest to the paleo-shoreline: more proximal), changing downdip (further away from the paleo-shoreline: distal) to more fine-grained, glauconitic quartzose silts and clays. The lower Eocene Manasquan Fm. commonly consists of a yellow-green to olive-green, massive calcareous clay/silts to very fine sands [Olsson and Wise, 1987]. A thick kaolinite-rich, clayey layer with rare or without glauconite is intercalated between the Vincentown and Manasquan lithologies [Olsson and Wise, 1987; Gibson *et al.*, 1993, 2000]. This unit has been assigned to the Marlboro Fm. [Kopp *et al.*, 2009] and is the regional expression of the PETM [e.g., Gibson *et al.*, 1993; 2000; Cramer *et al.*, 1999; Harris *et al.*, 2010]. The clays are unconformably overlain by glauconitic clays to clayey silts of the Manasquan Fm. [Olsson and Wise, 1987]. We refer to this unconformable contact as the main unconformity. Additional stratigraphic gaps may be recognized at sharp lithologic changes or by prominent changes in nannoplankton composition [e.g., Cramer *et al.*, 1999; Gibbs *et al.*, 2006a].

[6] Ocean Drilling Program Leg 174AX recovered continuous cores of uppermost Paleocene and lowermost Eocene deposits on the NJCP (including Ancora and Bass River) [Miller *et al.*, 1998, 1999]. The United States Geological Survey drilled a core at a more landward location at Clayton, supplemented by the Wilson Lake core, ~5 km to the East (Figure 1). High resolution foraminiferal and bulk stable isotope [Cramer *et al.*, 1999; Zachos *et al.*, 2006; Sluijs *et al.*, 2007b; John *et al.*, 2008], nannofloral [Bybell and Self-Trail, 1997; Gibbs *et al.*, 2006a, 2006b], organic biomarker and dinocyst [Zachos *et al.*, 2006; Sluijs and Brinkhuis, 2009] records have been generated, but benthic foraminiferal assemblages have been studied at much lower resolution only [Gibson *et al.*, 1993; Harris *et al.*, 2010]. Clayton and Wilson Lake are closest to the paleo-shoreline (updip), at shallowest depths, with Ancora intermediate and Bass River furthest offshore (downdip). The differences in water depth at the deepest Bass River and shallowest Clayton site has been estimated at about 40–70 m [Olsson and Wise, 1987; Harris *et al.*, 2010], with depths interpreted between 50 and 200 m (middle to outer shelf) at all sites over the studied interval.

**Table 1.** Previously Published Estimates of PETM Completeness, Total Amount of Time Represented in the Site and Average Sedimentation Rates

	Reference <sup>a</sup>	Completeness PETM Interval	Duration PETM Interval	Average Sedimentation Rate
Wilson Lake	2 and 3	relatively complete	<210–220 kyr	8.4 cm/kyr
	4	relatively complete	<170 kyr	8.4 cm/kyr
	5	incomplete	±100kyr	13.7 cm/kyr
	6	incomplete	±38.5–48 kyr	26–32.5 cm/kyr <sup>b</sup>
	this study	incomplete	±72 kyr	19 cm/kyr
Bass River	1	complete	not applicable	1.93 cm/kyr <sup>c</sup>
	2 and 3	relatively complete	<210–220 kyr	6.7 cm/kyr
	4	incomplete	±100–120 kyr	8–10 cm/kyr
	5	relatively complete	<200 kyr	5.1 cm/kyr
	6	incomplete	±100–120 kyr	8–10 cm/kyr
	this study	incomplete	±105 kyr	9.8 cm/kyr

<sup>a</sup>References: 1: *Cramer et al.* [1999]; 2: *Gibbs et al.* [2006a]; 3: *Gibbs et al.* [2006b]; 4: *Sluijs et al.* [2007b]; 5: *John et al.* [2008]; 6: *Sluijs and Brinkhuis*, [2009].

<sup>b</sup>Linear extrapolation of Bass River sedimentation rates based on *Sluijs and Brinkhuis* [2009].

<sup>c</sup>Uppermost Paleocene sedimentation rate extrapolated to the PETM sequence.

[7] Environmental changes during the PETM are clearly expressed in the Marlboro Fm. clays in New Jersey (NJ). Sea surface temperatures rose by up to 8°C during the early PETM, with peak temperatures in excess of 33°C [*Zachos et al.*, 2006]. The CIE is pronounced in bulk and foraminiferal carbonate as well as in organic matter of dinocysts [*Cramer et al.*, 1999; *Kent et al.*, 2003; *Zachos et al.*, 2006; *Sluijs et al.*, 2007b; *John et al.*, 2008]. Distinctive nannoplankton (*Discoaster araneus* and *D. anartios*) and dinocyst (*Apectodinium augustum*) associations are present [*Bybell and Self-Trail*, 1997; *Cramer et al.*, 1999; *Gibbs et al.*, 2006a; *Sluijs and Brinkhuis*, 2009]. Benthic foraminifera show assemblage changes [*Olsson and Wise*, 1987; *Gibson et al.*, 1993; *Gibson and Bybell*, 1994; *Cramer et al.*, 1999; *Harris et al.*, 2010], though considerably fewer taxa became extinct than in the deep sea. Benthic assemblages suggest that during the PETM dysoxic conditions prevailed, while primary productivity was high, as supported by nannofossil and dinocyst data [*Gibbs et al.*, 2006b; *Sluijs and Brinkhuis*, 2009]. The PETM silty clays were deposited at high sediment accumulation and carbon burial rates [*John et al.*, 2008], and high concentrations of kaolinite indicate intensive weathering [*Gibson et al.*, 1993; 2000; *Cramer et al.*, 1999]. The environmental setting during the PETM allowed proliferation as well as preservation of magnetotactic bacteria, resulting in abundant magnetofossils [*Lippert and Zachos*, 2007; *Kopp et al.*, 2009; *Dickens*, 2008]. Proliferation and preservation of magnetofossils indicate dysoxic (but not anoxic to euxinic) conditions within the sediments, at high iron bioavailability [*Dickens*, 2008].

[8] Despite the many studies, it remains unclear exactly which parts of the PETM are represented in the sediments at different locations due to unconformities and condensed intervals, so that estimates of the average sedimentation rates and completeness of the PETM diverge widely (Table 1). Biostratigraphic correlations between Wilson Lake and Bass River have been proposed [*Sluijs and Brinkhuis*, 2009; *Stassen et al.*, 2012a], but no integrated correlation between all four sites is available, and high-resolution benthic foraminiferal evidence has thus far not been included. We present high-resolution stable isotope records from the benthic species *Cibicidoides succedens/alleni* and *Anomalinoidea acutus*, addressing potential taxonomic inconsistencies between

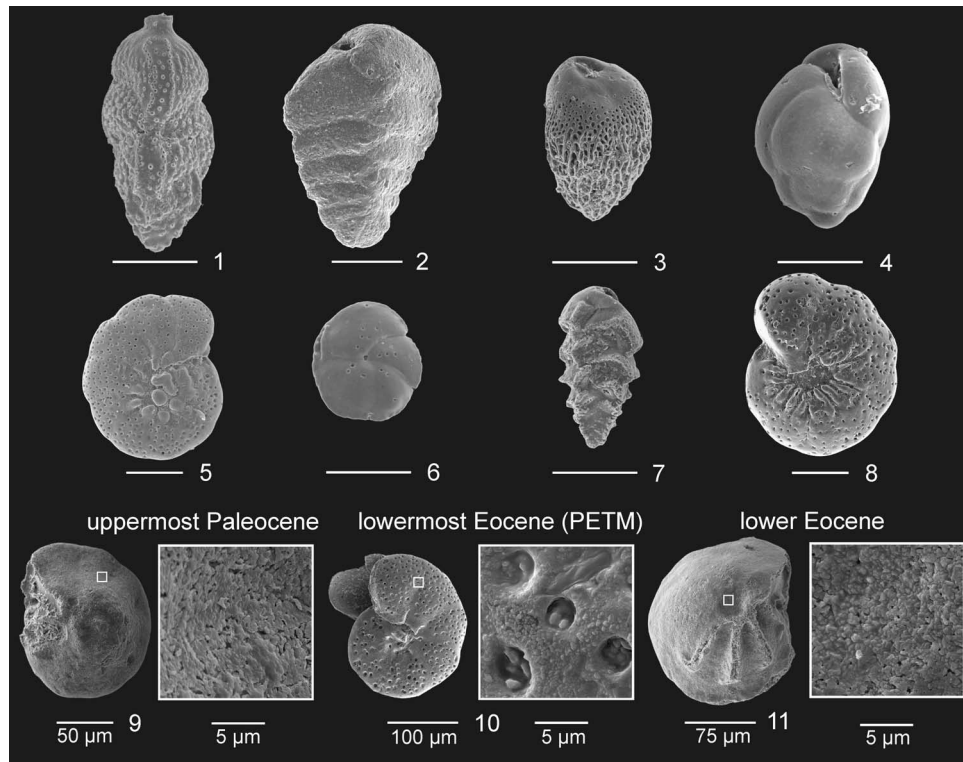
earlier studies [*Stassen et al.*, 2009], and compile distribution patterns of characteristic benthic foraminiferal taxa.

### 3. Methods

[9] Samples from Bass River and Wilson Lake were disintegrated in distilled water and washed over a >63 μm sieve, and the weight percentage of the sand-size fraction determined. Foraminiferal specimens in many samples in the Marlboro Fm. have a translucent, glassy test, whereas those from the sandy intervals are non-translucent (Figure 2). Foraminiferal tests from the PETM interval generally do not show significant etching or dissolution, although some specimens are filled with iron oxides or pyrite. We added data to the bulk carbonate isotope record of Clayton, with analyses performed at the University of California, Santa Cruz. Foraminiferal isotope records of the Wilson Lake and Bass River sites are based on multiple (10–20) specimens of single foraminiferal species from the 125–630 μm size fraction (*A. acutus* and *C. allenisuccedens*, Figure 3). Rare specimens in the virtually barren Paleocene are severely corroded and were not analyzed. Stable carbon isotopes of benthic foraminifera were measured with a Thermo DeltaPlus Advantage mass spectrometer at the Earth System Center for Stable Isotopic Studies (ESCSIS, Yale Institute for Biospheric Studies, Yale University). All values are given in δ<sup>13</sup>C notation relative to PeeDee belemnite standard (PDB) (see auxiliary material).<sup>1</sup>

[10] We integrated our data with published biostratigraphic (calcareous nannoplankton: NP) and lithostratigraphic (weight percentage > 63 μm: wt% and unconformities) data. Subdivision into nannoplankton (sub)zones follows *Aubry* [1996, 1999], a refinement of *Martini* [1971]. The lowest occurrence of *Discoaster araneus* or *D. anartios* in standard Zone NP9 allows the separation into subzones NP9a and NP9b. Zone NP10 is subdivided using the total ranges of *Tribrachiatus digitalis* and *T. contortus* (Subzones NP10b and 10d respectively) and the partial range of *T. bramlettei* (Subzones NP10a and NP10c). The NP9b–NP10 zonal boundary (lowest occurrence of *T. bramlettei*) is placed within the Marlboro Fm. by *Gibson et al.* [1993], *Gibbs et al.* [2006a], and L. Bybell (personal communication, 2010), differing from the placement

<sup>1</sup>Auxiliary materials are available at <ftp://ftp.agu.org/apend/pa/2012PA002323>.



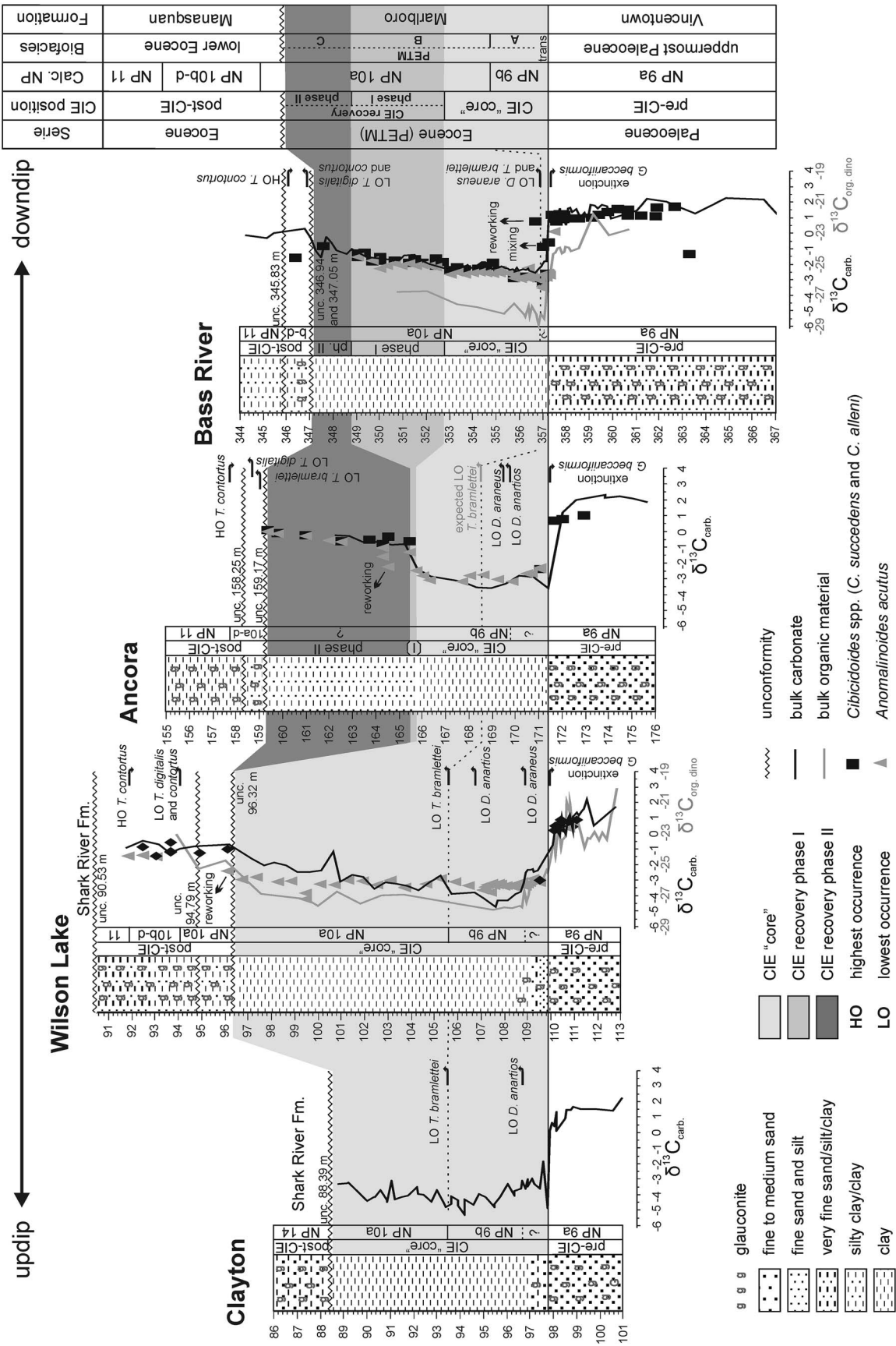
**Figure 2.** Scanning electron microscope images of important benthic foraminiferal taxa and details of the wall structures, displaying generally well preserved tests with minor test dissolution, recrystallization or overgrowth (WL: Wilson Lake, BR: Bass River, scale bar represents 100  $\mu\text{m}$ , unless otherwise indicated). 1: *Pseudovigenerina wilcoxensis* (WL 100.31 m), 2: *Spiroplectinella laevis* (BR 349.32 m), 3: *Bulimina callahani* (WL 96.04 m), 4: *Turrilina brevispira* (WL 95.43 m), 5: *Anomalinoidea acutus* (WL 108.84 m), 6: *Pulsiphonina prima* (WL 108.84 m), 7: *Tappanina selmensis* (106.41 m), 8: *Gavelinella beccariiformis* (BR 357.33 m), 9: *Cibicidoides alleni* (WL 92.69 m), 10: *Anomalinoidea acutus* (WL 107.02 m), 11: *Cibicidoides alleni* (110.14 m).

at the base of the sandy Manasquan Fm. [Miller *et al.*, 1999; Aubry *et al.*, 2000], probably due to different taxonomic concepts. We used the lowermost reported occurrence within the Marlboro Fm. as the zonal marker, in accordance with the observed range in deep-sea sequences [e.g., Agnini *et al.*, 2007]. Multiple lower Eocene unconformities are reported in the NJCP (Table 2), based on biostratigraphic gaps, erosional or burrowed surfaces, and variations in lithology.

[11] We determined planktic-benthic foraminiferal ratios (P/B), a paleodepth proxy [van der Zwaan *et al.*, 1990; Gibson, 1989; Gibson *et al.*, 1993], and relative abundances of selected benthic foraminiferal taxa (Table 3 and Figure 4) on representative splits (size fraction > 63  $\mu\text{m}$  at Wilson Lake and Bass River). In samples with abundant terrigenous grains, flotation in zinc chloride solution was used to concentrate foraminifera [Semensatto and Dias-Brito, 2007]. We used published data for Clayton (size fraction > 63  $\mu\text{m}$  [Gibson *et al.*, 1993]) and Ancora (size fraction > 100  $\mu\text{m}$  [Cramer *et al.*, 2000; Harris *et al.*, 2010]). No detailed benthic foraminiferal taxonomy of the NJ Paleocene/Eocene species has been published since Olsson [1960], and taxonomic use appears to be not consistent. For example, the distribution of *Bulimina aspero-aculeata* Brotzen 1948 at Bass River [Cramer *et al.*, 2000] mimics the distribution of the morphologically similar *Bulimina callahani* Galloway and Morrey 1931 in our taxonomy, and these are probably the same

species. A PETM bloom of *Spiroplectinella* and *Pseudovigenerina* species has been described at all sites, yet under different names (e.g., Gibson *et al.* [1993]: *Spiroplectammina wilcoxensis* Cushman and Ponton 1932; *Pseudovigenerina wilcoxensis* Cushman and Ponton 1932; we recognized *Spiroplectammina laevis* (Roemer 1841) and *Pseudovigenerina wilcoxensis*; Cramer *et al.* [2000] and Harris *et al.* [2010]: *Spiroplectammina plummerae* Cushman 1948 and *Pseudovigenerina cuneata* (Brotzen 1948). Detailed benthic foraminiferal taxonomy is not the topic here, and species belonging to these genera were combined to avoid taxonomic discrepancies.

[12] Benthic foraminifera in the PETM can be combined in biogroups (suites of taxa with shared distribution patterns) [Stassen *et al.*, 2012a, 2012b]. The abundance patterns of the main representatives of the three biogroups (Table 3; species shown in Figure 2) were used to define ecostratigraphic intervals, called biofacies (Figure 4). Benthic foraminiferal assemblage composition is mainly determined by the interplay between organic flux, bottom and pore water oxygenation, and species competition [van der Zwaan *et al.*, 1999; Murray, 2006; Jorissen *et al.*, 2007]. The main parameter in muddy, shallow water environments, where food-supply is much higher than in the deep-sea, is the availability of oxygen [van der Zwaan *et al.*, 1999; Jorissen *et al.*, 2007]. Species in biogroup 1 (*Tappanina selmensis*, *Pulsiphonina*



**Figure 3.** Biostratigraphy based on the lowest and highest occurrences of nannoplankton marker taxa (data sources in Table 2). Isotopic correlation along the NJCP is mainly based on benthic foraminiferal isotope records, in addition to dinocyst and bulk carbonate isotope records (data sources in Table 2). CIE subdivision corresponds to the chronology of *Röhrl et al.* [2007] and the lowest occurrence of *Trifarctiatus bramlettei* (zonal marker of the NP9-NP10 boundary).

**Table 2.** Sources of Chemostratigraphic, Biostratigraphic, Lithostratigraphic and Ecostratigraphic Data<sup>a</sup>

	Clayton	Wilson Lake	Ancora	Bass River
>63 $\mu\text{m}$ weight percentage	2	1	3	1
Nannoplankton	2	4, 5	9	4, 12
Planktic-benthic ratio	2	1	11	1
Benthic foraminifera	2	1	10, 11	1
$\delta^{13}\text{C}_{\text{bulk, carbonate}}$	1, 3	6	3	13
$\delta^{13}\text{C}_{\text{foraminifera}}$	–	1	16	1, 14
$\delta^{13}\text{C}_{\text{dinocyst, organic}}$	–	7	–	7
Unconformities	2	1, 8	9	14, 15

<sup>a</sup>References: 1: this study; 2: Gibson *et al.* [1993]; 3: Kent *et al.* [2003]; 4: Gibbs *et al.* [2006a]; 5: L. Bybell, unpublished data, 2010; 6: Zachos *et al.* [2006]; 7: Sluijs *et al.* [2007b]; 8: Gibbs *et al.* [2006b]; 9: Miller *et al.* [1999]; 10: Cramer *et al.* [2000]; 11: Harris *et al.* [2010]; 12: Aubry *et al.* [2000]; 13: John *et al.* [2008]; 14: Cramer *et al.* [1999]; 15: Miller *et al.* [1998]; 16: Cramer and Kent [2005].

*prima* and *Anomalinoidea acutus*) and biogroup 2 (*Pseudovigierina* and *Spiroplectinella*) have been generally regarded as indicators of dysoxic conditions and an elevated organic flux (eutrophic to mesotrophic conditions) [e.g., Olsson and Wise, 1987; Gibson *et al.*, 1993; Gibson and Bybell, 1994]. Most of these probably had an endobenthic lifestyle, because their morphology is similar to that of extant taxa under such conditions [Corliss, 1991; Buzas *et al.*, 1993; Jorissen *et al.*, 2007]. *Tappanina selmensis* is a shelf species, but became globally abundant in the deep ocean just after the benthic foraminiferal extinction [Thomas, 1998], and may be an opportunistic taxon, occurring at high abundance in severely disturbed or stressed environments [e.g., Gibson *et al.*, 1993].

## 4. Results

### 4.1. Lithostratigraphy and Biostratigraphy

[13] The Uppermost Paleocene Vincentown Fm. contains abundant quartz and glauconite at all sites (Figure 4), and is assigned to nannofossil Subzone NP9a (presence of *D. multiradiatus*, absence of PETM marker taxa). The benthic foraminiferal species *Gavelinella beccariiiformis* is a common cosmopolitan Paleocene benthic foraminiferal species ranging from the deep-sea into oligotrophic deep shelf settings [e.g., Widmark and Speijer, 1997], its extinction coincides with the onset of the CIE [Thomas, 1998, 2007]. It is common at the NJ sites, except for the shallowest (Clayton), and its highest occurrence (as globally) coincides with the onset of the CIE (Figure 3). Sporadically occurring specimens of *G. beccariiiformis* above the CIE-onset have pre-CIE isotope values [Cramer *et al.*, 1999], and are considered reworked, rather than representing a small diachronous range in shallow water ecosystems.

[14] At Clayton and Wilson Lake, there is a gradual transition from the typical Vincentown Fm. to the Marlboro Fm., where sediments become finer grained (from fine/medium sand, toward very fine silty sand up to clay up-core) and glauconite gradually disappears. This transitional interval has not been described at Ancora, and is absent in the most downdip Bass River (Figure 4). The fine-grained glauconite-poor clayey lithology is similar to the clays of the Marlboro Fm. in Maryland and Virginia, suggesting a large areal extent of this unit [Kopp *et al.*, 2009], which in Maryland, however,

represents a more marginal environment than in NJ [Gibson *et al.*, 1993].

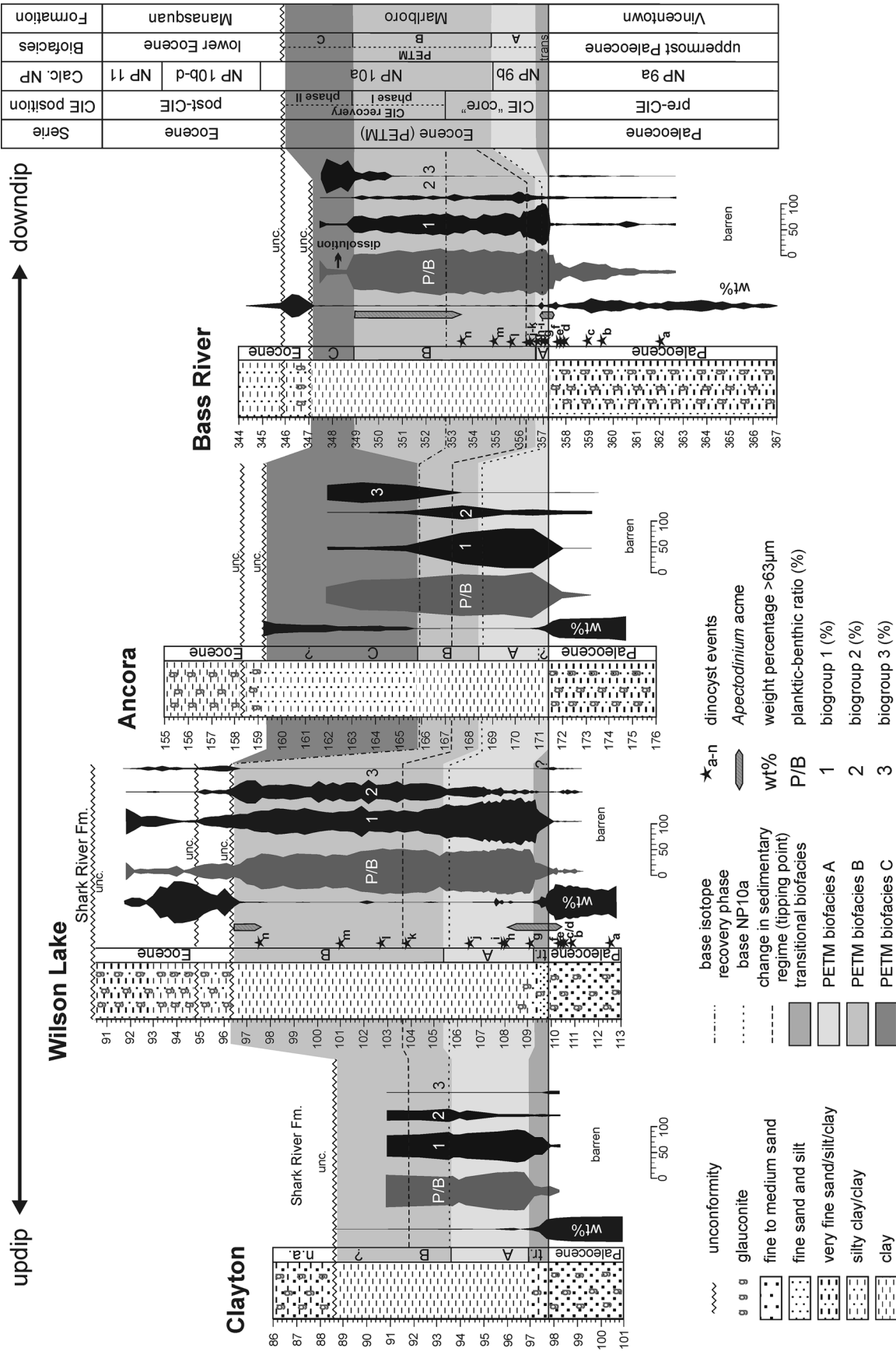
[15] Exact stratigraphic assignment of the transition to a nannoplankton zone is not possible because of severe dissolution of carbonate [Gibbs *et al.*, 2006a]. The clayey part of the Marlboro Fm. contains PETM nannofossil excursion taxa (either *Discoaster araneus* or *D. anartios*), allowing assignment to Subzone NP9b, which spans several m at Clayton and Wilson Lake, as marked by the lowermost occurrence of *Tribrachiatus bramlettei* (base NP10a). At Bass River, the lowest occurrences of *D. araneus* and *T. bramlettei* coincide, indicating minor discontinuities or strong condensation of the NP9b Subzone, in agreement with the lack of transitional sediments [John *et al.*, 2008].

[16] At Ancora, the fine-grained interval below the main unconformity consists of a clayey lower and a siltier upper part (Figure 4), which has been attributed respectively to the Marlboro and Manasquan Fm., with the placement of an unconformity at 165.51 m [e.g., Kopp *et al.*, 2009]. We disagree and assign the entire interval to the Marlboro Fm., based on benthic foraminiferal data (see 4.3) and the lack of glauconite, a common feature of the Manasquan Fm. PETM marker nannofossil taxa (*D. araneus* and *D. anartios*) are most abundant in the lower PETM interval (164.47–169.47 m), and only *D. araneus* occurs occasionally higher up, with its highest occurrence at 159.29 m [Miller *et al.*, 1999]. The Ancora site thus contains a 6.34 m thick lithologic unit not encountered in the other sites, entirely assigned to Subzone NP9b, with the lowest occurrence of *T. bramlettei* above the main unconformity (Figure 3). This suggests that Ancora contains the most expanded NP9b interval, based on low-resolution data [Miller *et al.*, 1999]. More detailed studies may show the lowermost occurrence of *T. bramlettei* closer to base of the clay unit at Ancora.

[17] The main unconformity is at the base of the lower Eocene glauconite-bearing Manasquan Fm. At most sites, Subzone NP10a is truncated at the top by the main unconformity and is most expanded at Bass River (Figure 3). A second major unconformity lies at the base of the Shark River Fm. (NP14), which directly overlies the Manasquan Fm. (Wilson Lake) or the Marlboro Fm. (Clayton) in the updip sites. Nannofossil distributions at Wilson Lake indicate a relatively expanded upper NP10a up to lower NP11 sequence (pers. comm., Laurel Bybell), whereas this interval is more condensed and incomplete downdip (Figure 3). At Bass River the sediment just below the main unconformity contains more coarse-grained particles (quartz and glauconite), but correlation with the silty interval at Ancora is uncertain,

**Table 3.** Summary of the Biotic Indicators Applied in This Study, Including Paleoecologic Characteristics and Biogroup Species

Bioindicator	Species	Paleoecologic Signal
high P/B ratio	not applicable	high surface production and/or poor bottom water oxygenation
biogroup 1	<i>Tappanina selmensis</i> <i>Anomalinoidea acutus</i> <i>Pulsiphonina prima</i>	continuously stressed dysoxic bottom conditions (river outflow assemblages)
biogroup 2	<i>Spiroplectinella</i> species <i>Pseudovigierina</i> species	periodically stressed dysoxic bottom conditions
biogroup 3	<i>Turritina brevispira</i> <i>Bulimina callahani</i>	eutrophic oxic bottom conditions



**Figure 4.** Litho- and ecostratigraphic correlation along the NJCP based on the planktic-benthic ratio, coarse fraction and distribution patterns of several characteristic taxa, combined in biogroups (data source in Table 2, biogroups composition: Table 3). *Apectodinium* abundance increases and dinocyst events (a-n) after *Sluijs and Brinkhuis* [2009].

and intense bioturbation across the unconformity may be the cause of the coarse grain size.

#### 4.2. Carbon Isotope Stratigraphy

[18] The CIE in NJ is observed in bulk carbonate, dinocyst organic carbon, and in benthic and planktic foraminifera (Figure 3) [Cramer *et al.*, 1999; Kent *et al.*, 2003; Zachos *et al.*, 2006; Shuijs *et al.*, 2007b; John *et al.*, 2008]. The onset of the CIE is defined by a rapid decrease in carbon isotope values, followed by sustained low  $\delta^{13}\text{C}$  values. This sharp negative shift is abrupt in the downdip sites (Ancora: 171.6–171.3 m; Bass River 357.4–357.3 m) [Cramer *et al.*, 1999; Kent *et al.*, 2003], without transitional values in bulk carbonate, dinocyst and benthic foraminiferal records (Figure 3), and without transitional  $\delta^{13}\text{C}$  values in single-specimen planktic foraminifera records at Bass River [Zachos *et al.*, 2007; John *et al.*, 2008]. The scatter in values around the P-E boundary at Bass River (357.4–356.6 m) is probably due to reworking: *Cibicidoides* is very rare in the basal PETM, so that isotope values probably reflect a mixture of reworked uppermost Paleocene and in situ lowermost Eocene specimens within this condensed interval [Cramer *et al.*, 1999; Stassen *et al.*, 2012a].

[19] At Wilson Lake, dinocyst  $\delta^{13}\text{C}$  values place the onset of the CIE between 109.9 and 109.8 m, followed by a gradual excursion toward lowest values from 109.8 to 108.7 m [Shuijs *et al.*, 2007b]. The dinocyst values within the CIE “core” range between  $-27$  and  $-28\text{‰}$  at Wilson Lake, and  $\sim 1\text{‰}$  lower at Bass River. The bulk carbonate and benthic foraminiferal isotope records are interrupted by a carbonate-free dissolution interval at the base of the CIE (Wilson Lake 109.9–109.5 m). The bulk carbonate record displays a rapid but gradual upwards decline in the initial stages of the CIE, which may reflect a declining effect of sediment mixing. The bulk carbonate isotope record of Clayton places the onset between 97.7 and 97.8 m [Gibson *et al.*, 1993], likewise interrupted by a thin carbonate-poor interval, and followed by decreasing values up to 95.5 m. At all sites, the benthic foraminiferal  $\delta^{13}\text{C}$  values remain stable in the succeeding expanded interval (Figure 3) with sustained low  $\delta^{13}\text{C}$  values (means at Wilson Lake:  $-3.3\text{‰}$ ; Ancora:  $-2.8\text{‰}$ ; Bass River:  $-2.4\text{‰}$ ). Planktic values (*Acarinina*) are  $\sim -0.6\text{‰}$  at Bass River, with a similar average but larger variability at Wilson Lake [Zachos *et al.*, 2006; John *et al.*, 2008].

[20] In the more updip sites (Clayton and Wilson Lake), the interval of sustained low benthic foraminiferal  $\delta^{13}\text{C}$  values continues up to the erosional contact at the main unconformity. At Wilson Lake, *Cibicidoides* spp. above this level yield post-CIE values. Immediately above the main unconformity (96.32 m), *A. acutus* is less well-preserved than *Cibicidoides*, suggesting reworking. At Bass River, a recovery trend starts at 352.8 m and continues up to the main unconformity. At Ancora, a marked initial isotope recovery occurs between 165.7 and 165.4 m, followed by a more gradual recovery, truncated by the main unconformity. The 2‰ range in isotope values between 164 and 164.5 m is attributed to reworking, although no changes in biota or lithology are observed. The two intervals in which  $\delta^{13}\text{C}$  increases upwards at Ancora are tentatively correlated to recovery phase I and II of the CIE, based on the distribution of PETM biofacies C (see 4.3), which suggests that the

uppermost part of the PETM at Bass River correlates with the second positive isotope trend (recovery phase II) at Ancora.

[21] In contrast to foraminiferal and dinocysts values, bulk carbonate  $\delta^{13}\text{C}$  in the “core” of the CIE varies strongly between sites, with much lower values at the updip sites than at the downdip sites (Figure 3). Bulk carbonate values at Bass River are on average  $-2.1\text{‰}$  [John *et al.*, 2008] (slightly lower than the benthic foraminiferal values), values at Ancora about  $-3.2\text{‰}$  [Kent *et al.*, 2003], whereas at Clayton and Wilson Lake values were as low as  $-5.3$  to  $-4.0\text{‰}$  in the lower part of the CIE, increasing to  $-3$  to  $-2\text{‰}$  higher up [Kent *et al.*, 2003; Zachos *et al.*, 2006]. The bulk carbonate  $\delta^{13}\text{C}$  values at South Dover Bridge in Maryland are about  $-3.0\text{‰}$ , close to these at Ancora [Self-Trail *et al.*, 2012].

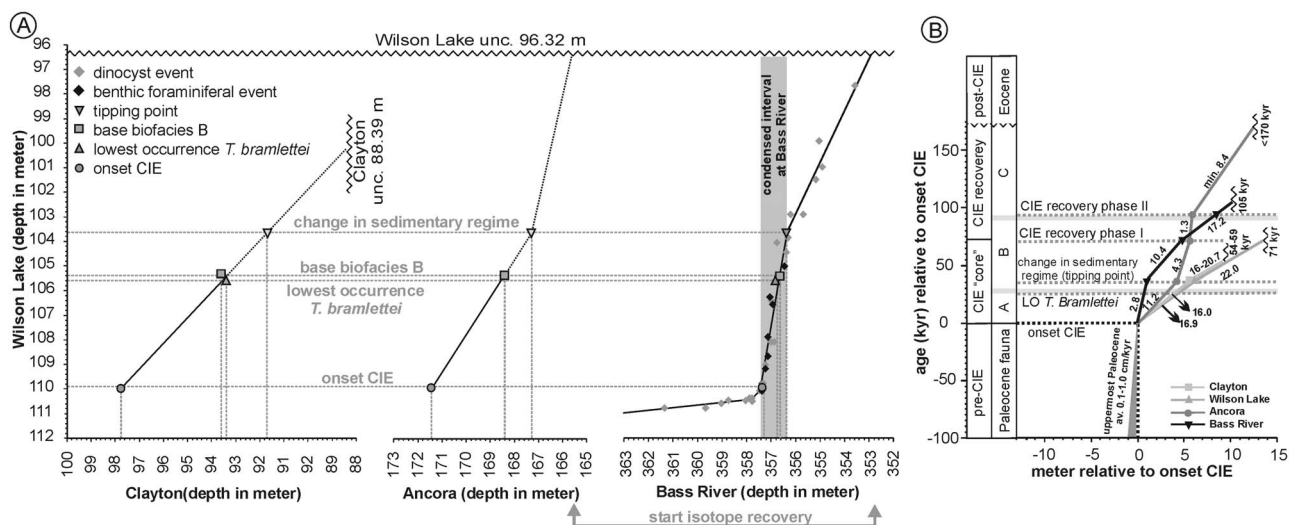
[22] Carbon isotope patterns during the CIE vary between the studied sites (Figure 3), with the largest differences in the bulk carbonate record, which may have been influenced by varying proportions of carbonate particles (nanofossils, planktic and benthic foraminiferal fragments), and/or diagenesis (see below). The dinocyst isotope record is based on total assemblages and may have an environmental/assemblage bias as fluctuating dinocyst abundances indicate dynamic surface water conditions during the PETM [Shuijs and Brinkhuis, 2009]. Planktic foraminifera show similar mean values at different sites, but the variability is larger at Wilson Lake [Zachos *et al.*, 2006; John *et al.*, 2008]. Benthic foraminiferal assemblages are relatively stable during the CIE at all sites, suggesting less variable bottom water conditions. Therefore we consider benthic foraminiferal  $\delta^{13}\text{C}$  records based on single-species best suitable for first order correlations, although some local heterogeneity in bottom waters cannot be excluded. The analyzed species (*A. acutus* and *C. succedens/alleni*) have both epibenthic morphologies, with almost identical isotope values in samples where they co-occur, indicating mutual vital effects.

[23] The Marlboro Fm. can thus be subdivided into three chemostratigraphic units (Figure 3). Wilson Lake contains the most expanded basal part of the PETM and all sites contain a zone of sustained low  $\delta^{13}\text{C}$  values, truncated at the top in the updip sites. The Bass River and Ancora isotope records indicate that a higher part of the PETM interval is present at the deeper sites, in agreement with interpretation of the dinocyst record at Bass River [Shuijs and Brinkhuis, 2009]. The PETM interval below the regional main unconformity includes here the top part of the CIE “core” and parts of the recovery phase (Bass River: recovery phase I and in part phase II; Ancora: mainly recovery phase II). Recovery phase II is expanded at Ancora, though incomplete, encompassing the silty interval and probably represents the youngest preserved part of the PETM in the NJCP.

#### 4.3. Ecostratigraphy

[24] Uppermost Paleocene (Vincentown) sediments are characterized by low P/B (planktic/benthic foraminiferal) values, as are post-PETM Eocene (Manasquan) sediments. In the uppermost Paleocene *G. beccariiiformis* is common and the species typical of the three PETM biogroups have minor occurrences (Figure 4). In the updip sections there are several intervals with very few to no calcareous foraminifera, whereas calcareous microfossils are more common at coeval levels at Bass River [Stassen *et al.*, 2012a]. The Marlboro clays deposited during the PETM contrast sharply





**Figure 5.** (a) Correlation between the NJCP sites (depth-depth plot) based on biotic, sedimentary and isotope events. The correlation between Wilson Lake and Bass River is based on benthic foraminiferal and dinocyst events [modified after Stassen *et al.*, 2012a]. (b) Depth-time plot based on the relative position and relative cyclostratigraphic ages of the CIE tie points [Röhl *et al.*, 2007] and lowest occurrence of *Tribrachiatus bramlettei* [Agnini *et al.*, 2007].

in foraminiferal content with both older and younger deposits, and having persistent high P/B values and sequential blooms of benthic foraminiferal species of the three PETM biogroups. Because the P/B is constantly high throughout the PETM, we used the distribution of the benthic foraminiferal biogroups to subdivide the interval into three biofacies (A, B and C).

[25] The biotic turnover between the uppermost Paleocene and the base of the Eocene is gradual in the updip sites (Clayton and Wilson Lake), corresponding to the gradual lithologic transition (Figure 4), where P/B values increase rapidly but gradually up-section, as does the abundance of biogroup 1 species [Stassen *et al.*, 2012a]. A thin barren dissolution zone occurs within the transitional interval [Gibbs *et al.*, 2006a; Zachos *et al.*, 2006; Stassen *et al.*, 2012a]. Sample resolution is insufficient to resolve the rate of turnover at Ancora, but the transition is abrupt at Bass River [Stassen *et al.*, 2012a], where there is also a rather abrupt lithologic contact and nannofossil Subzone NP9b is very thin. Extremely high abundances of planktic and benthic foraminiferal occur just above the contact [Stassen *et al.*, 2012a], suggesting that the transition into the PETM is punctuated by a short unconformity or condensed interval [John *et al.*, 2008]. The latter may be more probable, because the succession of regional biotic events at Bass River resembles that at the more expanded Wilson Lake [Sluijs and Brinkhuis, 2009; Stassen *et al.*, 2012a].

[26] At all sites, the lower part of the Marlboro Fm. contains biofacies A, dominated by biogroup 1 species *P. prima*, *A. acutus* and *T. selmensis*, and characterized by the lowest diversity. This biofacies is indicative of highly stressed seafloor conditions, probably related to low-oxygen conditions, as also recognized at other NJCP sites [Olsson and Wise, 1987; Gibson *et al.*, 1993; Harris *et al.*, 2010]. The presence of a relatively thick biofacies A interval at Ancora indicates that there is no condensed basal interval. The top of biofacies A is defined by the increase in abundances of

biogroup 2 species (*Spiroplectinella* and *Pseudouigerina*) and is closely associated with the NP9b-NP10a zonal boundary. Species of biogroup 1 remain common within the following biofacies. Biogroup 2 species are more common at the shallower sites and probably represents periodically rather than continually stressed, dysoxic bottom conditions [Stassen *et al.*, 2012a]. The abundances of the biogroups 1 and 2 become stable at 103.65 m at Wilson Lake, a level coeval with the top of the relatively condensed interval at Bass River (Figure 5). At Clayton and Wilson Lake, the top of biofacies B is truncated by the main unconformity.

[27] Downdip (Ancora and Bass River), biofacies B is succeeded by biofacies C, characterized by higher abundance of biogroup 3 species (Figure 4). This biofacies occurs in the top part of the clayey interval at Bass River and in the silty interval at Ancora, truncated at both sites by the main unconformity. Its continuation in the siltier unit at Ancora suggests that there is an expanded upper PETM interval at Ancora, not present at the other sites. Biofacies C indicates the re-establishment of more oxic bottom conditions, although at persistent eutrophic conditions. Planktic foraminifera and all three PETM biogroups persist above the main unconformity at Wilson Lake, where the overlying lower Eocene glauconitic sandy clays of the Manasquan Fm. contain a fauna comparable to that in the uppermost Paleocene, and the PETM biogroups are only minor components.

#### 4.4. Integrated Stratigraphy

[28] The NJCP Paleocene/Eocene transition can be subdivided into cross-gradient, correlatable intervals integrating biostratigraphic, lithostratigraphic, ecostratigraphic and chemostratigraphic data. Detailed dinocyst and benthic foraminiferal distribution patterns of the Wilson Lake and Bass River records [Sluijs and Brinkhuis, 2009; Stassen *et al.*, 2012a] indicate a consistent sequence of stratigraphic events during the PETM (Figures 4 and 5). Both sites contain similar, diverse nannoplankton assemblages [Gibbs *et al.*, 2006a] and

a simultaneous dispersal of *T. bramlettei* along the NJ shelf is expected. An uncertainty in the chemostratigraphic correlation is caused by the potential heterogeneity of the isotope composition of dissolved inorganic carbon (DIC) in shelf settings (see 5.1). The start of the isotope recovery trend, as observed in all carbon-bearing marine compounds, at Bass River (352.8 m), does not correlate to any level below the main unconformity at Wilson Lake (Figure 5). The absence of a benthic biofacies shift at this level indicates that there was no major change in seafloor conditions. Therefore, the isotopic shift at 352.8 m at Bass River probably reflects a genuine shift in the DIC pool, thus can be used for cross-gradient chronostratigraphic correlation.

[29] Regional litho- and ecostratigraphic correlations commonly are diachronous, and benthic foraminiferal distributions are generally related to paleodepth. Studies along a depth gradient thus are used to trace lateral facies shifts over time [e.g., Olsson and Wise, 1987]. As an example, during sea level rise, facies expressed at deeper sites move inland, to the locations of shallower sites [e.g., Harris et al., 2010]. Yet during the PETM, correlations based on trends in foraminiferal abundances may contain significant regional (eco)stratigraphic information, because of the widespread effect of the PETM on lithologic and biotic expressions. As an example, the termination of biofacies A closely corresponds to the NP9-NP10 zonal boundary. The characteristic PETM biofacies have not been observed anywhere along the eastern U.S. coastal Plain before the PETM [Gibson and Bybell, 1994], and rapidly became established over the full depth range of sites. Under these circumstances, one cannot use benthic biofacies to estimate paleodepth changes during PETM, because other factors than depth (sedimentary, environmental and taphonomic) were determinant. Estimates of changes in paleodepth using the foraminiferal data at face value (i.e., 50–70 m) [Harris et al., 2010] therefore tend to be higher than realistic estimates of sea level rise over the maximum of 10–20 kyr during the onset of PETM global warming, at a time where polar ice sheets were probably small (i.e., 20–30 m) [Shuijs et al., 2008].

[30] In conclusion, the vertical succession and lateral distribution of biofacies and lithofacies over the studied interval dominantly reflect the regional environmental evolution over time, i.e., the effects of the PETM global warming along the entire shelf gradient, and do not dominantly relate to water depth-related lateral facies shifts. The fact that the PETM environmental changes were the main determinant of the observed changes in lithology, benthic foraminiferal compositions and isotope values implies that the NJCP PETM interval can be subdivided into cross-gradient correlatable intervals, because the chemo-, bio- and isotope stratigraphic events were regionally synchronous at the time resolution of our study. The apparent minor inconsistencies between sites are likely due to differences in sample resolution of the studies.

[31] It had been suggested that the thickness of the Marlboro Fm. is a function of the proximity to the coast line, with thickest PETM intervals on the inner to middle shelf, thinning offshore [John et al., 2008]. We argue that sedimentation rates were indeed higher in the shallower locations, but that the total thickness of the PETM strata is a function of combined accumulation rates and post-PETM erosion, i.e., the extent of the main unconformity. We thus see much more

heterogeneity in the preserved part of the sedimentary record than formerly thought and conclude that the NJ PETM record remains incomplete, lacking specifically the later part of the recovery of the CIE in the shallowest settings.

#### 4.5. Age Model

[32] In the NJCP PETM sedimentation rates were high (Table 1), but accurate estimates are difficult due to multiple unconformities and condensed levels. Correlatable tie points in  $\delta^{13}\text{C}$  records in marine and terrestrial settings are well-calibrated [e.g., Zachos et al., 2005; Röhl et al., 2007]. The exact numerical chronology of the tie points remains debatable, in part because of discrepancies between cyclostratigraphic [Röhl et al., 2007] and  $^3\text{He}$  [Murphy et al., 2010] based age models, but with our basic correlation model these numerical ages can be adjusted as age models further develop. The uppermost Paleocene average sedimentation rates were calculated on the basis of the lowest occurrence of *Discoaster multiradiatus* (base NP9a, Clayton [Gibson et al., 1993]; Ancora [Miller et al., 1999]) or the base of magnetostratigraphic C24r (Bass River [Cramer et al., 1999]). Wilson Lake is linearly correlated with Bass River (Figure 5). The datum points within the PETM are the bases of recovery phases I and II, respectively 71.25 and 94.23 kyr relative to the base of the CIE at a duration of 170 kyr [Röhl et al., 2007], supplemented by the lowest occurrence of *T. bramlettei* (marker of base NP10a), with an age of 26 kyr after the onset of the PETM [Agnini et al., 2007].

[33] At all sites, the boundary between biofacies A and B occurs below a major change in sedimentary regime, with a condensed interval at Bass River [Stassen et al., 2012a]. The change in depositional regime is expressed in the clay mineralogy and corresponds to a decline in kaolinite abundance at Clayton and Bass River [Gibson et al., 1993; Cramer et al., 1999]. The start of the change in sedimentation pattern (tipping point) is extrapolated to all sites, using a site-to-site correlation between Wilson Lake (103.65 m) and Bass River (356.3 m), or a linear extrapolation at Clayton (91.79 m) and Ancora (167.24 m). The latter is based on the position of the onset of the CIE and the base of biofacies B relative to the position at Wilson Lake (Figure 5). Constant sedimentation rates are used within this basal part of the PETM and the lowermost occurrence of *T. bramlettei* is used to estimate average sediment accumulation rates. The positions of the CIE tie points allow an estimate of average accumulation rates in the succeeding intervals, and a minimum sedimentation rate for recovery phase II at Ancora (Figure 5).

[34] Average accumulation rates of the uppermost Paleocene Vincentown Fm. were on the order of 0.1–1 cm/kyr (Clayton: 0.7, Wilson Lake: 0.1, Ancora: 0.8 and Bass River: 1.0 cm/kyr). The rates are better constrained at Wilson Lake than at Clayton, where it is probably overestimated. The very low average sedimentation rates increased by at least an order of magnitude during the PETM, but with considerable differences between sites (Figure 5). Estimated sedimentation rates within the basal part of the PETM range from 2.8 (Bass River) to 16.9 cm/kyr (Wilson Lake), and within the CIE “core” from 4.3 cm/kyr at Ancora to 20.0 cm/kyr at Wilson Lake. The relative position of the top of the Marlboro Fm. at Wilson Lake is correlated to a coeval level at Bass River (Figure 5), indicating an average accumulation rate of 22.0 cm/kyr within the uppermost PETM at Wilson Lake

(103.65–96.32 m). The calculated total duration of the part of the PETM reflected in the sediments at Wilson Lake is thus 71 kyr, implying an almost complete CIE “core.” If the lowermost PETM accumulation rates remained constant at Clayton, or more likely increased at a rate similar to that in nearby Wilson Lake (respectively 16.0 and 20.7 cm/kyr), a time interval equivalent to approximately 54 to 59 kyr is represented at Clayton. Accumulation rates during the PETM peak warming are thus considerably higher from the level of change in sedimentary regime (tipping point) onwards up to the base of recovery phase I (except for Ancora), ranging from 4.3 (Ancora) to 22.0 cm/kyr (Wilson Lake).

[35] Average sedimentation rates decreased during recovery phase I at Ancora (1.3 cm/kyr), whereas they almost doubled at Bass River (17.2 cm/kyr). If the sedimentation rates during recovery phase II at Bass River remained constant, the total duration of the preserved PETM interval would be in the order of 105 kyr, in line with a tentative estimate of 100 kyr based on the cyclic fluctuations in abundance of *Apectodinium* [Sluijs *et al.*, 2007b]. Sedimentation rates during the terminal phase of the recovery at Ancora were at least 8.4 cm/kyr (max. total duration of the CIE: 170 kyr) as it is not known how much of the top part has been eroded. Sedimentation rates of the Manasquan Fm. (lower Eocene) cannot be calculated because of the lack of reliable data points, with thin sediment intervals bracketed by unconformities.

## 5. Discussion

### 5.1. Magnitude of the Carbon Isotope Excursion

[36] The PETM global warming has been attributed to the rapid emission of a large amount of isotopically light carbon into the ocean-atmosphere system. The magnitude of the CIE is used to estimate the amount of carbon compounds emitted, making assumptions regarding its isotopic composition [Pagani *et al.*, 2006]. However, there are large discrepancies between the magnitude of the CIE in marine and in terrestrial settings, in different types of marine carbonate records (bulk carbonate, planktic and benthic foraminiferal carbonate), and in marine carbonate records from different locations [Bowen *et al.*, 2006; McInerney and Wing, 2011]. The marine carbonate record of the CIE may be incomplete (thus an under-estimate) at many deep-sea locations because of ocean acidification and widespread dissolution of seafloor carbonates, and bulk organic carbon records may be over-estimates due to the impact of changes in vegetation. Data from Walvis Ridge (SE Atlantic) and Tanzania suggest that the globally averaged CIE may have ranged between  $-3.5$  to  $-5.1\text{‰}$  [McCarren *et al.*, 2008; Handley *et al.*, 2008]. Benthic single-species foraminiferal carbon isotope values at our sites (Figures 3 and 6), show a CIE of about  $-4\text{‰}$ , as do planktic foraminifera (*Acarinina* species), i.e., within this global range.

[37] Coastal regions, especially during eutrophication, commonly have considerably lower  $\delta^{13}\text{C}$  values (by several ‰) in DIC and thus in foraminiferal carbonate [Thomas *et al.*, 2000; Diz *et al.*, 2009] than open ocean, due to the oxidation of land-derived or marine-bloom-related organic matter. This effect needs consideration when examining isotope records along continental shelves under the influence of river discharge [Dickens, 2011], especially those derived

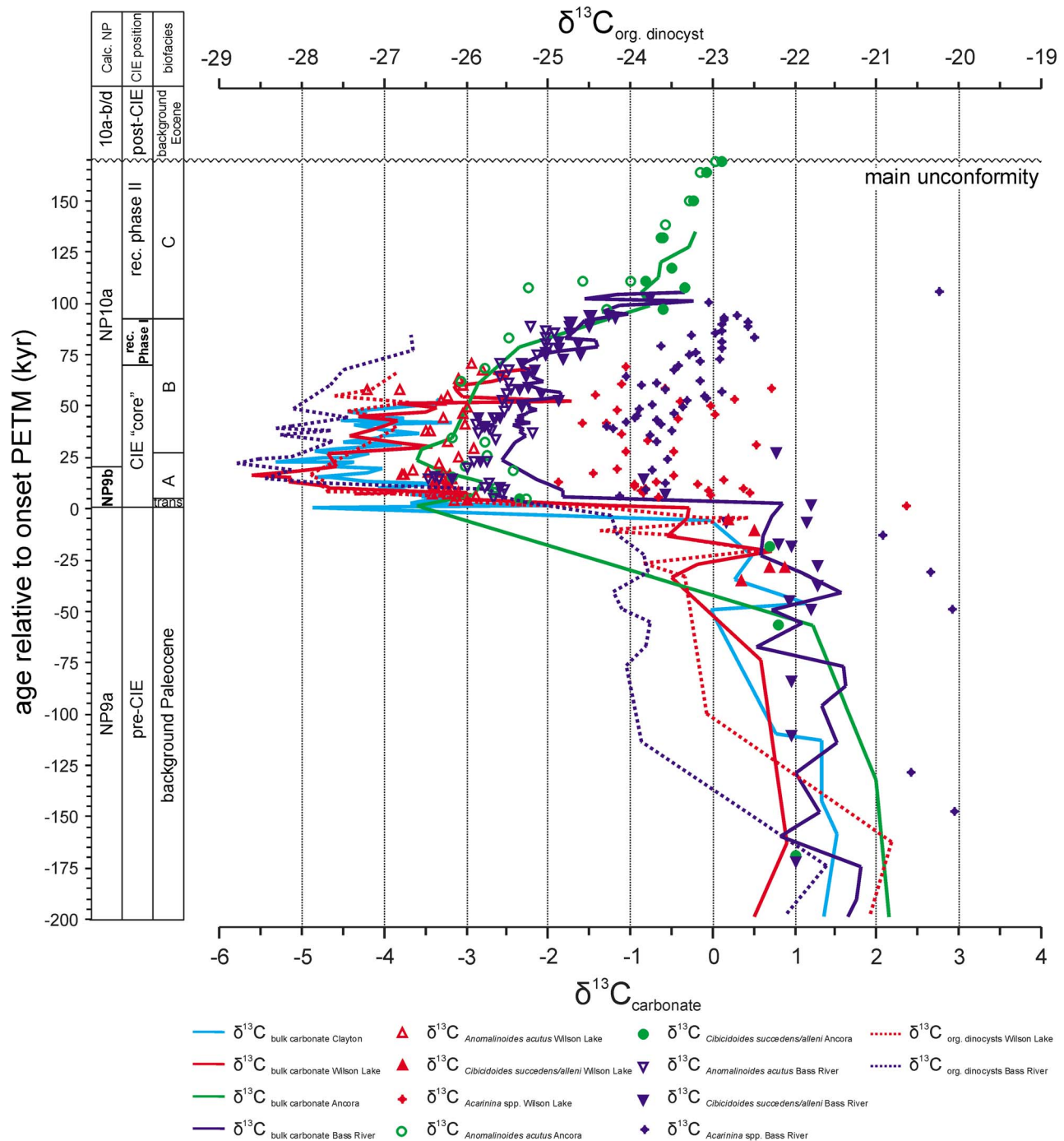
from surface-dwelling organisms. The benthic values before and during the PETM tend to be slightly lower in the more proximal settings (Wilson Lake) than in the more distal settings (Bass River), but this difference is only about 0.5–0.7‰, and may be the expected coastal gradient. The greater variability in planktic foraminiferal  $\delta^{13}\text{C}$  at Wilson Lake might likewise be due to the coastal effect. The CIE in dinoflagellate  $\delta^{13}\text{C}$  values (Figure 6) is of similar magnitude ( $\sim -4\text{‰}$ ) as the foraminiferal CIE, but values overall are somewhat less negative at Wilson Lake than at Bass River, i.e., in the opposite direction as expected from the coastal trend. These values are derived from a mixture of species, however, thus might incorporate the effect of differences in species composition [Sluijs and Brinkhuis, 2009].

[38] We cannot fully explain the large differences between the magnitudes of the CIE in bulk carbonate  $\delta^{13}\text{C}$  at our and other sites in the Salisbury Embayment [e.g., *Self-Trail et al.*, 2012]. Values at Wilson Lake are lower by up to 3‰ during the early part of the CIE than at Bass River, with values for Ancora intermediate (Figure 6). If these values reflected the value of DIC, we should see similar differences in magnitude in the CIE in the foraminiferal and dinocysts records, but they are absent. It seems unlikely that the minor differences in the taxonomic composition of calcareous nannoplankton [Bybell and Self-Trail, 1995; Gibbs *et al.*, 2006a], the main component of bulk carbonate, would cause such a major difference [Stoll, 2005]. The differences between the magnitude of the bulk carbonate CIE are thus not easily explained, but might be due to diagenetic effects on fine fraction carbonate in low-carbonate samples, as observed in some deep-sea sections [Bralower *et al.*, 1998; Zachos *et al.*, 2005]. We thus conclude that benthic foraminiferal isotope records probably are not significantly affected by local influences, and can be used to correlate the different sites, allowing improved correlations within the United States Coastal Plain, as well with the global record.

### 5.2. Preservation of Magnetofossils

[39] The Marlboro Fm. contains a varying mixture of detrital and biogenic magnetic particles. The high deposition rate during the PETM is associated with limited oxygenation at the seafloor, as expressed by the poorly diverse benthic foraminiferal assemblages (Figure 7). Magnetite-precipitating bacteria occur and are also preserved in this specific environment, resulting in an abundance increase of magnetofossils in the Marlboro Fm. [Dickens, 2008; Kopp *et al.*, 2009]. The observed regional, lateral trend in the magnetofossil-detrital ratio, with decreasing biogenic particles at the more distal sites, led to the assumption that suboxic conditions within the sediment mainly occurred in the inner to middle shelf, and extension of such conditions to the flank (Bass River) of the Salisbury Embayment was transient [Kopp *et al.*, 2009].

[40] We argue that the inferred lateral trends are not coeval, because the sediments deposited during time of dysoxic conditions are more condensed at Bass River, where magnetoparticles are mainly of detrital origin in the basal PETM, with magnetofossils higher up. This could indicate that there was no expanded suboxic zone within the sediment during the early stages of the PETM [Dickens, 2008]. Recovery phase I at Ancora and Bass River have magnetic particles of mixed biotic and terrigenous origin [Kopp *et al.*, 2009] indicative of higher periodic reoxygenation of bottom waters.

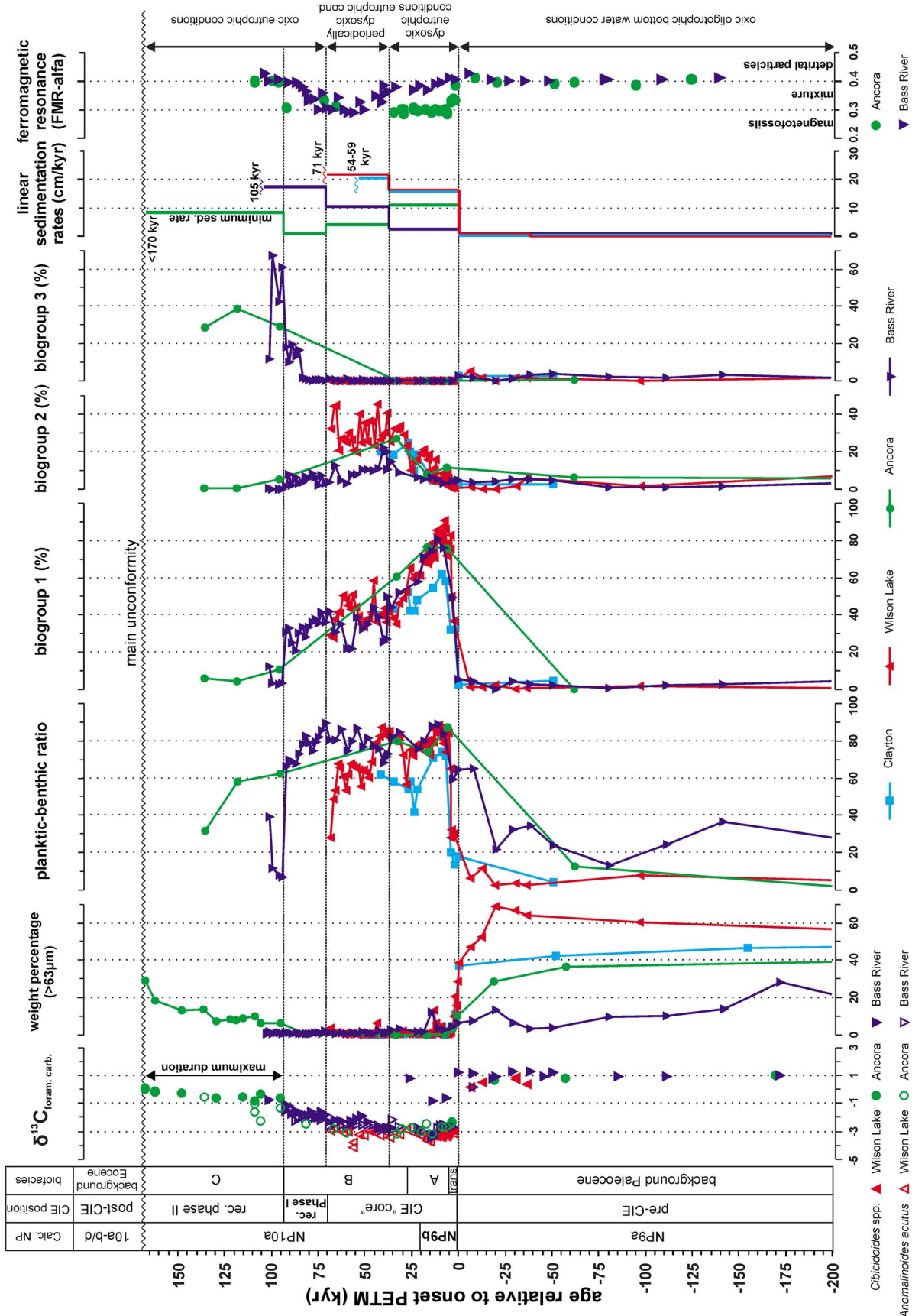


**Figure 6.** Comparison of the marine isotope records of the NJCP, plotted against the proposed age model (data from Kent et al. [2003], Cramer and Kent [2005], Zachos et al. [2006], Sluijs et al. [2007b], and John et al. [2008]).

This transition was not previously noticed at Ancora, because recovery phase I is relatively condensed. Recovery phase II contains fewer magnetic particles of mainly terrigenous origin [Kopp et al., 2009], indicating the non-generation or non-preservation of magnetofossils. The increase in grain size at Ancora (silty interval) indicates renewed sorting of the detrital input, probably due to intensified bottom currents, allowing blooms of biogroup 3 taxa. This gradual increasing oxygenation caused reduced stocks of magnetotactic bacteria

and their final disappearance when bottom water oxygenation improved during recovery phase II at Ancora and Bass River (Figure 7).

[41] Our proposed stratigraphic subdivision of the Marlboro Fm. in the NJCP into a CIE “core” and two recovery phases thus places constraints on the interpretation of the magnetic particle distribution in cores in NJ and further to the South, and requires re-evaluation of all stratigraphic correlations [e.g., Kopp et al., 2009]. We cannot address this



**Figure 7.** Proposed age model of the Paleocene/Eocene transition in the NJCP, emphasizing the time represented in the sediments at each site. Magnetic particles after origin [Kopp et al., 2009]. The Ancora record is plotted in maximum ages (complete CIE recovery).

because no detailed biotic data from the cores further to the south are available, but because of the higher accumulation rates in the more proximal settings, only partial preservation of the CIE “core” is anticipated in the southern part of the Salisbury Embayment.

### 5.3. Sedimentary Regime

[42] The lower Paleogene NJCP has been characterized as a starved siliciclastic marine shelf with persistently high sea level and low sediment supply [Browning *et al.*, 2008]. Rivers with outflow on the shelf, probably located further to the South (paleo-Potomac and paleo-Susquehanna Rivers), provided only a minor sediment load to the Salisbury Embayment. The sedimentation rates for these sediments are very low (<1 cm/kyr), as is common for regions with glauconite sedimentation. In this open marine shelf setting with a low sediment supply, fine-grained material may have been winnowed by currents. In such sedimentary regimes with relatively high primary productivity, strong dissolution of calcium carbonate macro- and microfossils may occur [Green *et al.*, 1993]. The lack of planktic foraminifera in the Vincentown Fm. and the relatively poor preservation of benthic foraminifera thus may at least in part be a taphonomic phenomenon, with the more easily dissolved planktic foraminifera most affected [Nguyen *et al.*, 2009].

[43] The PETM is accompanied by a radical shift in sedimentation patterns [Gibson *et al.*, 2000]. River outflow and mud-deposition is strongly influenced by climatic conditions [Eisma *et al.*, 1991]. Intensification of the hydrologic cycle associated with the PETM [Pierrehumbert, 2002] changed the local river outflow, leading to deposition of the thick clayey strata of the Marlboro Fm. at the proximal sites and lower accumulation rates in the outer shelf region (Bass River). The export of large amounts of suspended clay and silt onto the shelf may have resulted from the establishment of a tropical river-dominated shelf, due to increase of flow in the paleo-Potomac and paleo-Susquehanna rivers [Kopp *et al.*, 2009]. The strong increase in sedimentation rate may have led to greatly increased preservation of carbonate fossils. The high abundance of planktic foraminifera in sediments deposited during the PETM, in combination with their carbon isotope values and highly diverse nannoplankton assemblages [Gibbs *et al.*, 2006a] indicates open marine conditions without greatly reduced salinity. This suggests that river discharges occurred far from the studied sites, in agreement with the low abundance of terrestrial palynomorphs in the Marlboro Fm. in the NJCP although a 3–4 ppt drop in surface water salinity could have occurred [Zachos *et al.*, 2006]. The interplay of the increased river discharges and subsequent transport by shelf bottom currents may have resulted in the deposition of a mud belt in the northern Salisbury Embayment, with heterogeneity in sedimentation rates as observed in modern mud belts [Bianchi and Allison, 2009].

[44] Benthic foraminifera monitor fluvial impact on marine shelf ecosystems, even at large distances from the river outflow areas [van der Zwaan and Jorissen, 1991]. Due to the interaction of excess organic fluxes and periodic depletion of oxygen levels caused by its decomposition, organic matter can be captured within fine-grained sediments [Aller and Blair, 2006]. We interpret biogroup 1, with abundant *A. acutus*, *P. prima* and *T. selmensis*, as a river outflow assemblage, inhabiting an organic-rich mud belt as the direct

biotic response to the eutrophication and potential stratification of surface waters caused by large riverine inputs to which opportunistic taxa rapidly responded [e.g., van der Zwaan and Jorissen, 1991]. The dominance of this one biogroup in biofacies A during the first ~30 kyr of the PETM suggests continuous eutrophication of the entire shelf, possibly with salinity and/or thermal driven stratification of the water column and continual fluxes of suspended clay particles.

[45] The increase in sedimentation rates coincided with increasing abundance of low-salinity tolerant dinocysts [Sluijs and Brinkhuis, 2009] ~40 kyr after the onset of the PETM. From this level on, the abundances of both biogroup 1 and 2 taxa were fairly stable at Wilson Lake (Figure 7), possibly indicating recurrent flood events during monsoonal precipitation cycles [Zachos *et al.*, 2006] or high frequencies of tropical cyclone activity [Kopp *et al.*, 2009]. A brief intense wet season and prolonged dry season, creating a sparsely vegetated hinterland, has been proposed to explain the combination of high siliciclastic input and low terrestrial organic matter [Zachos *et al.*, 2006; Sluijs and Brinkhuis, 2009], but terrestrial and marine organic matter could have been partly remineralized after deposition [Aller and Blair, 2006]. The co-occurrence of two biogroups in biofacies B is the response to an establishment of a periodic (seasonal) pattern in stratification and dysoxia during the upper part of the CIE “core” and recovery phase I (Figure 7). During these periodic flood events, large amounts of suspended clay and fresh-water tolerant dinoflagellates were carried offshore, far from the coastline, and deposition occurred in an expanded mud belt on the entire NJ shelf.

[46] Environmental conditions ameliorated during recovery phase II, when eutrophic, yet oxic bottom water conditions became established as indicated by biofacies C. The reoxygenation of the water column and deposition of silt (Ancora) rather than clay indicate intensification of bottom currents (e.g., storm events), preventing continual stratification and allowing oxygenation at least locally. The continuation of high sedimentation rates indicates that riverine discharges persisted, but the river-dominated shelf evolved toward more storm-dominated conditions during recovery phase II.

### 5.4. Ocean Acidification

[47] The preservation of planktic and benthic foraminifera is excellent within the PETM interval, with glassy preservation, in contrast to the preservation in the uppermost Paleocene sands (non-translucent), especially at the shallower sites (Figure 2). The improved preservation may, at least in part, be due to the increased rates of sedimentation, but is surprising because it has been argued that ocean acidification was severe during the PETM, at the greatly increased atmospheric CO<sub>2</sub> levels [Zachos *et al.*, 2005]. Modeling suggested that severe dissolution during ocean acidification would have affected shallower waters strongly [Caldeira and Wickett, 2003]. The presence of ‘excursion taxa’ of calcareous nannofossils has been proposed to be a response to ocean acidification [Raffi *et al.*, 2009], and their occurrence at the NJ sites has been interpreted accordingly, although this is debated [Gibbs *et al.*, 2010; Self-Trail *et al.*, 2012]. A relatively slow rates of ocean acidification during the PETM (as measured in kyr compared to present-day acidification), however, could have resulted in minor decrease in carbonate saturation in surface waters and

much more severe undersaturation in the deeper ocean [Hönisch et al., 2012]. If this is correct, it could explain why carbonate preservation in shelf settings is better than in the deep ocean [Zachos et al., 2005], underscoring the importance of studying the effects of global warming in different settings in order to understand global impacts. If ocean acidification was an important cause of the deep-sea benthic foraminiferal extinction, less severe dissolution at shallower depths could explain the much less severe extinction among surface calcifiers such as calcareous nannoplankton and planktic foraminifera, as well as the much less severe extinction in shallow-water than in deep-sea benthic foraminifera.

## 6. Conclusions

[48] 1. Based on the integration of bio-, eco- and lithostratigraphic data from four NJCP sites, a regional correlation framework of the top of the Vincentown Fm. and the Marlboro Fm. is proposed and used to develop an age model for these shelf sequences.

[49] 2. The major climatic and environmental changes during the PETM determined general patterns in lithology, benthic foraminiferal isotope records and foraminiferal distribution across the studied shelf transect, which can be used for a stratigraphic subdivision of the PETM interval in the NJCP.

[50] 3. None of the studied sites along the NJ transect contains a complete coverage of the PETM because widespread but regionally variable unconformities have removed parts of the sequences to a variable degree. The early stages of the PETM are best captured at Clayton and Wilson Lake, where a transitional fauna and lithology is recorded. The Bass River and Ancora isotope records contain the higher parts of the CIE with (partial) preservation of recovery phases I and II below the regional main unconformity.

[51] 4. During the PETM, the NJ shelf changed from a sediment-starved setting to a tropical, river-dominated mudbelt system, probably due to the increased activity of the hydrologic cycle. Benthic biofacies and distribution of magnetofossils reflect the associated environmental changes. The highly increased sedimentation rates facilitated excellent preservation of carbonate microfossils, in sharp contrast to the very poor preservation prior to the PETM.

[52] 5. The excellent preservation of carbonate suggests a lack of ocean acidification in shelf settings. This may have been an important factor in the low degree of extinction among shelf-dwelling benthic foraminifera, compared to the deep-sea.

[53] **Acknowledgments.** This research used samples and data of the United States Geological Survey (USGS) and ODP Leg 174AX, provided by the Integrated Ocean Drilling Program (IODP). Sampling of the Bass River core was assisted by Mimi Katz. Sampling of the Clayton and Wilson Lake core was assisted by Laurel Bybell, Jean Self-Trail, Leah Schneider and Tim Bralower. We thank Appy Sluijs and an anonymous reviewer for helpful comments and discussion. Financial support was provided by the KU Leuven Research Fund and the Research Foundation Flanders (FWO) to Robert P. Speijer and by the National Science Foundation (NSF) to Ellen Thomas (grant OCE-0903014).

## References

Agneri, C., E. Fornaciari, I. Raffi, D. Rio, U. Röhl, and T. Westerhold (2007), High-resolution nannofossil biochronology of middle Paleocene to early Eocene at ODP Site 1262: Implications for calcareous

nannoplankton evolution, *Mar. Micropaleontol.*, 64(3–4), 215–248, doi:10.1016/j.marmicro.2007.05.003.

Aller, R. C., and N. E. Blair (2006), Carbon remineralization in the Amazon-Guianas tropical mobile mudbelt: A sedimentary incinerator, *Cont. Shelf Res.*, 26(17–18), 2241–2259, doi:10.1016/j.csr.2006.07.016.

Aubry, M.-P. (1996), Towards an upper Paleocene-lower Eocene high resolution stratigraphy based on calcareous nannofossil stratigraphy, *Isr. J. Earth Sci.*, 44(4), 239–253.

Aubry, M.-P. (1999), Late Paleocene–early Eocene sedimentary history in western Cuba: Implications for the LPTM and for regional tectonic history, in *Lower Paleogene Biostratigraphy of Cuba.*, *Micropaleontology*, vol. 45, suppl. 2, edited by R. H. Fluegeman and M.-P. Aubry, pp. 5–18, Micropaleontology, New York, doi:10.2307/1486101.

Aubry, M. P., B. S. Cramer, K. G. Miller, J. D. Wright, D. V. Kent, and R. K. Olsson (2000), Late Paleocene event chronology: Unconformities, not diachrony, *Bull. Soc. Geol. Fr.*, 171(3), 367–378, doi:10.2113/171.3.367.

Aubry, M.-P., et al. (2007), The global standard stratotype-section and point (GSSP) for the base of the Eocene series in the Dababiya section (Egypt), *Episodes*, 30(4), 271–286.

Bianchi, T. S., and M. A. Allison (2009), Large-river delta-front estuaries as natural “recorders” of global environmental change, *Proc. Natl. Acad. Sci. U. S. A.*, 106(20), 8085–8092, doi:10.1073/pnas.0812878106.

Bowen, G. J., et al. (2006), Eocene hyperthermal event offers insight into greenhouse warming, *Eos Trans. AGU*, 87, 165, doi:10.1029/2006EO170002.

Bralower, T. J., D. J. Thomas, E. Thomas, and J. C. Zachos (1998), High-resolution records of the late Paleocene thermal maximum and circum-Caribbean volcanism: Is there a causal link? Reply, *Geology*, 26(7), 671.

Browning, J. V., K. G. Miller, P. J. Sugarman, M. A. Kominz, P. P. McLaughlin, A. A. Kulpecz, and M. D. Feigenson (2008), 100 Myr record of sequences, sedimentary facies and sea level change from Ocean Drilling Program onshore coreholes, US Mid-Atlantic coastal plain, *Basin Res.*, 20(2), 227–248, doi:10.1111/j.1365-2117.2008.00360.x.

Buzas, M. A., S. J. Culver, and F. J. Jorissen (1993), A statistical evaluation of the microhabitats of living (stained) infaunal benthic foraminifera, *Mar. Micropaleontol.*, 20(3–4), 311–320, doi:10.1016/0377-8398(93)90040-5.

Bybell, L. M., and J. M. Self-Trail (1995), Evolutionary, biostratigraphic and taxonomic study of calcareous nannofossils from a continuous Paleocene-Eocene boundary section in New Jersey, *U.S. Geol. Surv. Prof. Pap.*, 1554, 36pp.

Bybell, L. M., and J. M. Self-Trail (1997), Late Paleocene and early Eocene calcareous nannofossils from three boreholes in an onshore-offshore transect from New Jersey to the Atlantic continental rise *Proc. Ocean Drill. Program Sci. Results*, 150, 91–110.

Caldeira, K., and M. E. Wickett (2003), Anthropogenic carbon and ocean pH, *Nature*, 425(6956), 365, doi:10.1038/425365a.

Corliss, B. H. (1991), Morphology and microhabitat preferences of benthic foraminifera from the northwest Atlantic Ocean, *Mar. Micropaleontol.*, 17(3–4), 195–236, doi:10.1016/0377-8398(91)90014-W.

Cramer, B. S., and D. V. Kent (2005), Bolide summer: The Paleocene/Eocene thermal maximum as a response to an extraterrestrial trigger, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 224(1–3), 144–166, doi:10.1016/j.palaeo.2005.03.040.

Cramer, B. S., M.-P. Aubry, K. G. Miller, R. K. Olsson, J. D. Wright, and D. V. Kent (1999), An exceptional chronologic, isotopic, and clay mineralogical record of the latest Paleocene thermal maximum, Bass River, NJ, ODP 174AX, *Bull. Soc. Geol. Fr.*, 170(6), 883–897.

Cramer, B. S., K. G. Miller, J. D. Wright, M.-P. Aubry, and R. K. Olsson (2000), Neritic records of the late Paleocene thermal maximum from New Jersey, *GFF*, 122, 38–39, doi:10.1080/11035890001221038.

Crouch, E. M., C. Heilmann-Clausen, H. Brinkhuis, H. E. G. Morgans, K. M. Rogers, H. Egger, and B. Schmitz (2001), Global dinoflagellate event associated with the late Paleocene thermal maximum, *Geology*, 29(4), 315–318, doi:10.1130/0091-7613(2001)029<0315:GDEAWT>2.0.CO;2.

Cui, Y., L. R. Kump, A. J. Ridgwell, A. J. Charles, C. K. Junium, A. F. Diefendorf, K. H. Freeman, N. M. Urban, and I. C. Harding (2011), Slow release of fossil carbon during the Paleocene-Eocene thermal maximum, *Nat. Geosci.*, 4(7), 481–485, doi:10.1038/ngeo1179.

DeConto, R. M., S. Galeotti, M. Pagani, D. Tracy, K. Schaefer, T. J. Zhang, D. Pollard, and D. J. Beerling (2012), Past extreme warming events linked to massive carbon release from thawing permafrost, *Nature*, 484(7392), 87–91, doi:10.1038/nature10929.

Dickens, G. R. (2008), Palaeoclimate: The riddle of the clays, *Nat. Geosci.*, 1(2), 86–88, doi:10.1038/ngeo118.

Dickens, G. R. (2011), Down the rabbit hole: Toward appropriate discussion of methane release from gas hydrate systems during the Paleocene-

- Eocene thermal maximum and other past hyperthermal events, *Clim. Past*, 7(3), 831–846, doi:10.5194/cp-7-831-2011.
- Dickens, G. R., M. M. Castillo, and J. C. G. Walker (1997), A blast of gas in the latest Paleocene: Simulating first-order effects of massive dissociation of oceanic methane hydrate, *Geology*, 25(3), 259–262, doi:10.1130/0091-7613(1997)025<0259:ABOGIT>2.3.CO;2.
- Diz, P., F. J. Jorissen, G. J. Reichart, C. Poulain, F. Dehairs, E. Leorri, and Y. M. Paulet (2009), Interpretation of benthic foraminiferal stable isotopes in subtidal estuarine environments, *Biogeosciences*, 6(11), 2549–2560, doi:10.5194/bg-6-2549-2009.
- Dupuis, C., M.-P. Aubry, E. Steurbaut, W. A. Berggren, K. Ouda, R. Magioncalda, B. S. Cramer, D. V. Kent, R. P. Speijer, and C. Heilmann-Clausen (2003), The Dababiya Quarry section: Lithostratigraphy, clay mineralogy, geochemistry and paleontology, *Micropaleontology*, 49, 41–59, doi:10.2113/49.Suppl.1.41.
- Eisma, D., P. G. E. F. Augustinus, and C. Alexander (1991), Recent and subrecent changes in the dispersal of Amazon mud, *Neth. J. Sea Res.*, 28(3), 181–192, doi:10.1016/0077-7579(91)90016-T.
- Gibbs, S. J., P. R. Bown, J. A. Sessa, T. J. Bralower, and P. A. Wilson (2006a), Nannoplankton extinction and origination across the Paleocene-Eocene thermal maximum, *Science*, 314(5806), 1770–1773, doi:10.1126/science.1133902.
- Gibbs, S. J., T. J. Bralower, P. R. Bown, J. C. Zachos, and L. M. Bybell (2006b), Shelf and open-ocean calcareous phytoplankton assemblages across the Paleocene-Eocene thermal maximum: Implications for global productivity gradients, *Geology*, 34(4), 233–236, doi:10.1130/G22381.1.
- Gibbs, S. J., H. M. Stoll, P. R. Bown, and T. J. Bralower (2010), Ocean acidification and surface water carbonate production across the Paleocene-Eocene thermal maximum, *Earth Planet. Sci. Lett.*, 295(3–4), 583–592, doi:10.1016/j.epsl.2010.04.044.
- Gibson, T. G. (1989), Planktonic benthonic foraminiferal ratios: Modern patterns and Tertiary applicability, *Mar. Micropaleontol.*, 15(1–2), 29–52, doi:10.1016/0377-8398(89)90003-0.
- Gibson, T. G., and L. M. Bybell (1994), Sedimentary patterns across the Paleocene-Eocene boundary in the Atlantic and Gulf Coastal Plains of the United States, *Bull. Soc. Belge Geol.*, 103(3–4), 237–265.
- Gibson, T. G., L. M. Bybell, and J. P. Owens (1993), Latest Paleocene lithologic and biotic events in neritic deposits of southwestern New Jersey, *Paleoceanography*, 8(4), 495–514, doi:10.1029/93PA01367.
- Gibson, T. G., L. M. Bybell, and D. B. Mason (2000), Stratigraphic and climatic implications of clay mineral changes around the Paleocene/Eocene boundary of the northeastern US margin, *Sediment. Geol.*, 134(1–2), 65–92, doi:10.1016/S0037-0738(00)00014-2.
- Green, M. A., R. C. Aller, and J. Y. Aller (1993), Carbonate dissolution and temporal abundances of foraminifera in Long Island Sound sediments, *Limnol. Oceanogr.*, 38(2), 331–345, doi:10.4319/lo.1993.38.2.0331.
- Guasti, E., and R. P. Speijer (2007), The Paleocene-Eocene thermal maximum in Egypt and Jordan: An overview of the planktic foraminiferal record, in *Large Ecosystem Perturbations: Causes and Consequences*, edited by S. Monechi, R. Coccioni, and M. R. Rampino, *Spec. Pap. Geol. Soc. Am.*, 424, 53–67, doi:10.1130/2007.2424(03).
- Handley, L., P. N. Pearson, I. K. McMillan, and R. D. Pancost (2008), Large terrestrial and marine carbon and hydrogen isotope excursions in a new Paleocene/Eocene boundary section from Tanzania, *Earth Planet. Sci. Lett.*, 275(1–2), 17–25, doi:10.1016/j.epsl.2008.07.030.
- Harris, A. D., K. G. Miller, J. V. Browning, P. J. Sugarman, R. K. Olsson, B. S. Cramer, and J. D. Wright (2010), Integrated stratigraphic studies of Paleocene-lowermost Eocene sequences, New Jersey Coastal Plain: Evidence for glacioeustatic control, *Paleoceanography*, 25, PA3211, doi:10.1029/2009PA001800.
- Hönisch, B., et al. (2012), The geological record of ocean acidification, *Science*, 335(6072), 1058–1063, doi:10.1126/science.1208277.
- John, C. M., S. M. Bohaty, J. C. Zachos, A. Sluijs, S. Gibbs, H. Brinkhuis, and T. J. Bralower (2008), North American continental margin records of the Paleocene-Eocene thermal maximum: Implications for global carbon and hydrological cycling, *Paleoceanography*, 23, PA2217, doi:10.1029/2007PA001465.
- Jorissen, F. J., C. Fontanier, and E. Thomas (2007), Paleoceanographical proxies based on deep-sea benthic foraminiferal assemblage characteristics, in *Proxies in Late Cenozoic Paleoceanography: Part 2. Biological Tracers and Biomarkers*, edited by C. Hillaire-Marcel and A. de Vernal, pp. 263–325, Elsevier, Amsterdam, doi:10.1016/S1572-5480(07)01012-3.
- Kelly, D. C., T. J. Bralower, J. C. Zachos, I. P. Silva, and E. Thomas (1996), Rapid diversification of planktonic foraminifera in the tropical Pacific (ODP Site 865) during the late Paleocene thermal maximum, *Geology*, 24(5), 423–426, doi:10.1130/0091-7613(1996)024<0423:RDOPFI>2.3.CO;2.
- Kennett, J. P., and L. D. Stott (1991), Abrupt deep-sea warming, paleoceanographic changes and benthic extinctions at the end of the Paleocene, *Nature*, 353(6341), 225–229, doi:10.1038/353225a0.
- Kent, D. V., B. S. Cramer, L. Lanci, D. Wang, J. D. Wright, and R. Van der Voo (2003), A case for a comet impact trigger for the Paleocene/Eocene thermal maximum and carbon isotope excursion, *Earth Planet. Sci. Lett.*, 211(1–2), 13–26, doi:10.1016/S0012-821X(03)00188-2.
- Koch, P. L., J. C. Zachos, and P. D. Gingerich (1992), Correlation between isotope records in marine and continental carbon reservoirs near the Paleocene/Eocene boundary, *Nature*, 358(6384), 319–322, doi:10.1038/358319a0.
- Kopp, R. E., D. Schumann, T. D. Raub, D. S. Powars, L. V. Godfrey, N. L. Swanson-Hysell, A. C. Maloof, and H. Vali (2009), An Appalachian Amazon? Magnetofossil evidence for the development of a tropical river-like system in the mid-Atlantic United States during the Paleocene-Eocene thermal maximum, *Paleoceanography*, 24, PA4211, doi:10.1029/2009PA001783.
- Kurtz, A. C., L. R. Kump, M. A. Arthur, J. C. Zachos, and A. Paytan (2003), Early Cenozoic decoupling of the global carbon and sulfur cycles, *Paleoceanography*, 18(4), 1090, doi:10.1029/2003PA000908.
- Lippert, P. C., and J. C. Zachos (2007), A biogenic origin for anomalous fine-grained magnetic material at the Paleocene-Eocene boundary at Willson Lake, New Jersey, *Paleoceanography*, 22, PA4104, doi:10.1029/2007PA001471.
- Martini, E. (1971), Standard Tertiary and Quaternary calcareous nannoplankton zonation, in *Proceedings of the 2nd Planktonic Conference*, edited by A. Farinacci, pp. 739–785, Tecnoscienza, Rome.
- McCarren, H., E. Thomas, T. Hasegawa, U. Rohl, and J. C. Zachos (2008), Depth dependency of the Paleocene-Eocene carbon isotope excursion: Paired benthic and terrestrial biomarker records (Ocean Drilling Program Leg 208, Walvis Ridge), *Geochem. Geophys. Geosyst.*, 9, Q10008, doi:10.1029/2008GC002116.
- McInerney, F. A., and S. L. Wing (2011), The Paleocene-Eocene thermal maximum: A perturbation of carbon cycle, climate, and biosphere with implications for the future, *Annu. Rev. Earth Planet. Sci.*, 39, 489–516, doi:10.1146/annurev-earth-040610-133431.
- Miller, K. G. (1997), Coastal plain drilling and the New Jersey sea-level transect, *Proc. Ocean Drill. Program Sci. Results*, 150, 3–11.
- Miller, K. G., et al. (1998), Bass River site, *Proc. Ocean Drill. Program Init. Rep.*, 174, 5–43.
- Miller, K. G., et al. (1999), Ancora Site, *Proc. Ocean Drill. Program Init. Rep.*, 174A, supplement, 1–65.
- Murphy, B. H., K. A. Farley, and J. C. Zachos (2010), An extraterrestrial He-3-based timescale for the Paleocene-Eocene thermal maximum (PETM) from Walvis Ridge, IODP Site 1266, *Geochim. Cosmochim. Acta*, 74(17), 5098–5108, doi:10.1016/j.gca.2010.03.039.
- Murray, J. W. (2006), *Ecology and Applications of Benthic Foraminifera*, Cambridge Univ. Press, Cambridge, U. K., doi:10.1017/CBO9780511535529.
- Nguyen, T. M. P., M. R. Petrizzo, and R. P. Speijer (2009), Experimental dissolution of a fossil foraminiferal assemblage (Paleocene-Eocene thermal maximum, Dababiya, Egypt): Implications for paleoenvironmental reconstructions, *Mar. Micropaleontol.*, 73(3–4), 241–258, doi:10.1016/j.marmicro.2009.10.005.
- Olsson, R. K. (1960), Foraminifera of latest Cretaceous and earliest Tertiary age in the New Jersey Coastal Plain, *J. Paleontol.*, 34(1), 1–58.
- Olsson, R. K., and S. W. Wise (1987), Upper Paleocene to middle Eocene depositional sequences and hiatuses in the New Jersey Atlantic Margin, in *Timing and Depositional History of Eustatic Sequences: Constraints on Seismic Stratigraphy*, *Spec. Publ. 24*, edited by C. A. Ross and D. Haman, pp. 99–112, Cushman Found. for Foraminiferal Res., Houston, Tex.
- Pagani, M., K. Caldeira, D. Archer, and J. C. Zachos (2006), An ancient carbon mystery, *Science*, 314(5805), 1556–1557, doi:10.1126/science.1136110.
- Pierrehumbert, R. T. (2002), The hydrologic cycle in deep-time climate problems, *Nature*, 419(6903), 191–198, doi:10.1038/nature01088.
- Raffi, I., J. Backman, and H. Pälike (2005), Changes in calcareous nannofossil assemblages across the Paleocene/Eocene transition from the paleo-equatorial Pacific Ocean, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 226(1–2), 93–126, doi:10.1016/j.palaeo.2005.05.006.
- Raffi, I., J. Backman, J. C. Zachos, and A. Sluijs (2009), The response of calcareous nannofossil assemblages to the Paleocene Eocene thermal maximum at the Walvis Ridge in the South Atlantic, *Mar. Micropaleontol.*, 70(3–4), 201–212, doi:10.1016/j.marmicro.2008.12.005.
- Röhl, U., T. Westerhold, T. J. Bralower, and J. C. Zachos (2007), On the duration of the Paleocene-Eocene thermal maximum (PETM), *Geochem. Geophys. Geosyst.*, 8, Q12002, doi:10.1029/2007GC001784.
- Self-Trail, J. M., D. S. Powars, D. K. Watkins, and G. A. Wandless (2012), Calcareous nannofossil assemblage changes across the Paleocene-Eocene thermal maximum: Evidence from a shelf setting, *Mar. Micropaleontol.*, 92–93, 61–80, doi:10.1016/j.marmicro.2012.05.003.



- Semensatto, D. L., and D. Dias-Brito (2007), Alternative saline solutions to float foraminiferal tests, *J. Foraminiferal Res.*, 37(3), 265–269, doi:10.2113/gsjfr.37.3.265.
- Sluijs, A., and H. Brinkhuis (2009), A dynamic climate and ecosystem state during the Paleocene-Eocene thermal maximum: Inferences from dinoflagellate cyst assemblages on the New Jersey Shelf, *Biogeosciences*, 6(8), 1755–1781, doi:10.5194/bg-6-1755-2009.
- Sluijs, A., G. J. Bowen, H. Brinkhuis, L. J. Lourens, and E. Thomas (2007a), The Paleocene-Eocene thermal maximum super greenhouse: Biotic and geochemical signatures, age models and mechanisms of global change, in *Deep-Time Perspectives on Climate Change: Marrying the Signal From Computer Models and Biological Proxies*, edited by M. Williams et al., pp. 323–350, Micropalaeontol. Soc., London.
- Sluijs, A., H. Brinkhuis, S. Schouten, S. M. Bohaty, C. M. John, J. C. Zachos, G. J. Reichart, J. S. Sinninghe Damsté, E. M. Crouch, and G. R. Dickens (2007b), Environmental precursors to rapid light carbon injection at the Palaeocene/Eocene boundary, *Nature*, 450(7173), 1218–1221, doi:10.1038/nature06400.
- Sluijs, A., et al. (2008), Eustatic variations during the Paleocene-Eocene greenhouse world, *Paleoceanography*, 23, PA4216, doi:10.1029/2008PA001615.
- Sluijs, A., J. C. Zachos, and R. E. Zeebe (2012), Constraints on hyperthermals, *Nat. Geosci.*, 5(4), 230–231, doi:10.1038/ngeo1423.
- Speijer, R. P., C. Scheibner, P. Stassen, and A.-M. M. Morsi (2012), Response of marine ecosystems to deep-time global warming: A synthesis of biotic patterns across the Paleocene-Eocene thermal maximum (PETM), *Aust. J. Earth Sci.*, 150(1), 6–12.
- Stassen, P., E. Thomas, and R. P. Speijer (2009), Benthic foraminiferal isotope records across the PETM from the New Jersey Coastal Plain, in *Climatic and Biotic Events of the Paleogene*, *GNS Sci. Misc. Ser.*, vol. 18, edited by E. M. Crouch, C. P. Strong, and C. J. Hollis, pp. 135–137, Inst. of Geol. and Nucl. Sci., Lower Hutt, New Zealand.
- Stassen, P., E. Thomas, and R. P. Speijer (2012a), The progression of environmental changes during the onset of the Paleocene-Eocene thermal maximum (New Jersey Coastal Plain), *Aust. J. Earth Sci.*, 105(1), 169–178.
- Stassen, P., E. Thomas, and R. P. Speijer (2012b), Restructuring outer neritic foraminiferal assemblages in the aftermath of the Palaeocene-Eocene thermal maximum, *J. Micropalaeontol.*, 31, 89–93, doi:10.1144/0262-821X11-026.
- Stoll, H. M. (2005), Limited range of interspecific vital effects in coccolith stable isotopic records during the Paleocene-Eocene thermal maximum, *Paleoceanography*, 20, PA1007, doi:10.1029/2004PA001046.
- Svensen, H., S. Planke, A. Malthes-Sorensen, B. Jamtveit, R. Myklebust, T. R. Eidem, and S. S. Rey (2004), Release of methane from a volcanic basin as a mechanism for initial Eocene global warming, *Nature*, 429(6991), 542–545, doi:10.1038/nature02566.
- Thomas, E. (1998), Biogeography of the Late Paleocene benthic foraminiferal extinction, in *Late Paleocene-Early Eocene Climatic and Biotic Events in the Marine and Terrestrial Records*, edited by M.-P. Aubry, S. H. Lucas, and W. A. Berggren, pp. 214–243, Columbia Univ. Press, New York.
- Thomas, E. (2007), Cenozoic mass extinctions in the deep sea: What perturbs the largest habitat on Earth?, in *Large Ecosystem Perturbations: Causes and Consequences*, edited by S. Monechi, R. Coccioni, and M. R. Rampino, *Spec. Pap. Geol. Soc. Am.*, 424, 1–23, doi:10.1130/2007.2424(01).
- Thomas, E., and N. J. Shackleton (1996), The Paleocene-Eocene foraminiferal extinction and stable isotope anomalies, in *Correlation of the Early Paleogene in Northwest Europe*, edited by R. W. Knox, R. M. Corfield, and R. E. Dunay, *Geol. Soc. Spec. Publ.*, 101, 401–441, doi:10.1144/GSL.SP.1996.101.01.20.
- Thomas, E., T. Gapotchenko, J. C. Varekamp, E. L. Mccray, and M. R. B. ten Brink (2000), Benthic foraminifera and environmental changes in Long Island Sound, *J. Coastal Res.*, 16(3), 641–655.
- van der Zwaan, G. J., and F. J. Jorissen (1991), Biofacial patterns in river-induced shelf anoxia, in *Modern and Ancient Continental Shelf Anoxia*, edited by R. V. Tyson and T. H. Pearson, *Geol. Soc. Spec. Publ.*, 58, 65–82.
- van der Zwaan, G. J., F. J. Jorissen, and H. C. Destigter (1990), The depth dependency of planktonic benthic foraminiferal ratios: Constraints and applications, *Mar. Geol.*, 95(1), 1–16, doi:10.1016/0025-3227(90)90016-D.
- van der Zwaan, G. J., I. A. P. Duijnste, M. den Dulk, S. R. Ernst, N. T. Jannink, and T. J. Kouwenhoven (1999), Benthic foraminifera: Proxies or problems? A review of paleoecological concepts, *Earth Sci. Rev.*, 46(1–4), 213–236, doi:10.1016/S0012-8252(99)00011-2.
- Widmark, J. G. V., and R. P. Speijer (1997), Benthic foraminiferal ecomarker species of the terminal Cretaceous (late Maastrichtian) deep-sea Tethys, *Mar. Micropaleontol.*, 31(3–4), 135–155, doi:10.1016/S0377-8398(97)00008-X.
- Zachos, J. C., et al. (2005), Rapid acidification of the ocean during the Paleocene-Eocene thermal maximum, *Science*, 308(5728), 1611–1615, doi:10.1126/science.1109004.
- Zachos, J. C., S. Schouten, S. Bohaty, T. Quattlebaum, A. Sluijs, H. Brinkhuis, S. J. Gibbs, and T. J. Bralower (2006), Extreme warming of mid-latitude coastal ocean during the Paleocene-Eocene thermal maximum: Inferences from TEX86 and isotope data, *Geology*, 34(9), 737–740, doi:10.1130/G22522.1.
- Zachos, J. C., S. M. Bohaty, C. M. John, H. McCarren, D. C. Kelly, and T. Nielsen (2007), The Palaeocene-Eocene carbon isotope excursion: Constraints from individual shell planktonic foraminifer records, *Philos. Trans. R. Soc. A*, 365(1856), 1829–1842, doi:10.1098/rsta.2007.2045.
- Zachos, J. C., G. R. Dickens, and R. E. Zeebe (2008), An early Cenozoic perspective on greenhouse warming and carbon-cycle dynamics, *Nature*, 451(7176), 279–283, doi:10.1038/nature06588.