

## An exceptional chronologic, isotopic, and clay mineralogic record of the latest Paleocene thermal maximum, Bass River, NJ, ODP 174AX

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*Key words.* – LPTM, Chron C24r, Nannofossil biostratigraphy, Paleocene, Clay mineralogy,  $\delta^{13}\text{C}$ ,  $\delta^{18}\text{O}$ .

*Abstract.* – A thick, apparently continuous section recording events of the latest Paleocene thermal maximum in a neritic setting was drilled at Bass River State Forest, New Jersey as part of ODP Leg 174AX [Miller, Sugarman, Browning *et al.*, 1998]. Integrated nannofossil and magneto-stratigraphy provides a firm chronology supplemented by planktonic foraminiferal biostratigraphy. This chronologic study indicates that this neritic section rivals the best deep-sea sections in providing a complete record of late Paleocene climatic events. Carbon and oxygen isotopes measured on benthic foraminifera show a major (4.0‰ in carbon, 2.3‰ in oxygen) negative shift correlative with the global latest Paleocene carbon isotope excursion (CIE). A sharp increase in kaolinite content coincides with the isotope shift in the Bass River section, analogous to increases found in several other records. Carbon and oxygen isotopes remain low and kaolinite content remains high for the remainder of the depositional sequence above the CIE (32.5 ft, 9.9 m), which we estimate to represent 300-500 k.y. We interpret these data as indicative of an abrupt shift to a warmer and wetter climate along the North American mid-Atlantic coast, in concert with global events associated with the CIE.

### Un enregistrement sédimentaire exceptionnel (chronologie, isotopes et minéralogie) du « Latest Paleocene Thermal Maximum » : la coupe de Bass River, New Jersey, ODP Leg 174AX

*Mots clés.* – LPTM, Chron C24r, Biostratigraphie des nannofossiles, Paléocène, Minéralogie des argiles,  $\delta^{13}\text{C}$ ,  $\delta^{18}\text{O}$ .

*Résumé.* – La coupe de Bass River, New Jersey, offre un excellent enregistrement des événements liés au Late Paleocene Thermal Maximum (LPTM) que nous interprétons comme témoignant d'un changement climatique soudain vers des conditions chaudes et humides sur la côte mid-atlantique du continent Nord-Américain.

#### VERSION FRANÇAISE ABRÉGÉE

Une épaisse séquence, apparemment continue, carottée à Bass River State Forest, New Jersey, dans le cadre du Leg ODP 174AX [Miller, Sugarman, Browning *et al.*, 1998], a enregistré en milieu néritique les événements liés au Latest Paleocene Thermal Maximum (LPTM). L'interprétation stratigraphique est fondée sur l'intégration de données magnétostratigraphiques et biostratigraphiques (nannofossiles calcaires et foraminifères planctoniques). Nous montrons ainsi que l'enregistrement du LPTM en milieu néritique dans la coupe de Bass River est de qualité au moins égale à celui de coupes océaniques (DSDP et ODP). Les isotopes du carbone et de l'oxygène, mesurés sur les tests de foraminifères benthiques, montrent un shift majeur (4,0‰ : carbone, 2,3‰ : oxygène) corrélatif de l'excursion isotopique du carbone de la fin du Paléocène (CIE). Comme dans d'autres coupes, ce shift isotopique dans la coupe de Bass River coïncide avec une augmentation abrupte de la teneur en kaolinite. Ces deux événements se sont produits au cours d'un intervalle estimé à < 10 k.y. Les valeurs isotopiques de l'oxygène et du carbone restent faibles, et la teneur en kaolinite reste élevée dans l'intervalle de 32.5 ft (9.9 m) susjacent à l'excursion et représentant au moins 300-500 k.y. Nous interprétons ces données comme témoignant d'un changement climatique soudain vers des conditions chaudes et humides le long de la côte mid-atlantique du continent nord-américain, en liaison avec les événements globaux associés avec la CIE.

#### INTRODUCTION

The latest Paleocene thermal maximum (LPTM) is an apparently unique event in Cenozoic climatic and oceanographic history. It is characterized by a large (~ 4‰) negative carbon isotope excursion (CIE) lasting < 20 k.y. which has been recorded in both marine and terrestrial records [*e.g.*, Kennett and Stott 1990, 1991; Koch *et al.*, 1992; Sinha and Stott, 1993; Schmitz *et al.*, 1996; Stott *et al.*, 1996]. The carbon isotope excursion was accompanied by a negative excursion (1-3‰) in oxygen isotopes [Kennett and Stott, 1990, 1991; Pak and Miller, 1992; Lu and Keller, 1993; Bralower *et al.*, 1995; but see Aubry, 1998 for discussion], a benthic foraminiferal extinction event in deep

marine records [*e.g.*, Beckman, 1960; Tjalsma and Lohmann, 1983; Miller *et al.*, 1987a; Thomas, 1989, 1990a, b, 1991; Katz and Miller, 1991; Pak and Miller, 1992; see Thomas, 1998 for further references], and a mammalian turnover in terrestrial records, marking the Clarkforkian/Wasatchian boundary in North America and the base of the Sparnacian "stage" in Europe [Gingerich, 1989; Hooker, 1991, 1996; Koch *et al.*, 1992; Sinha and Stott, 1993; Maas *et al.*, 1995]. More recently it has been shown that certain short-lived and evolutionarily anomalous species of planktonic foraminifera (the so-called "excursion taxa" *Acarinina africana*, *Acarinina sibaiaensis*, and *Morozovella allisonensis*, Kelly *et al.* [1996]) have their lowest occurrence at the level of the CIE. The quantity of isotopically negative

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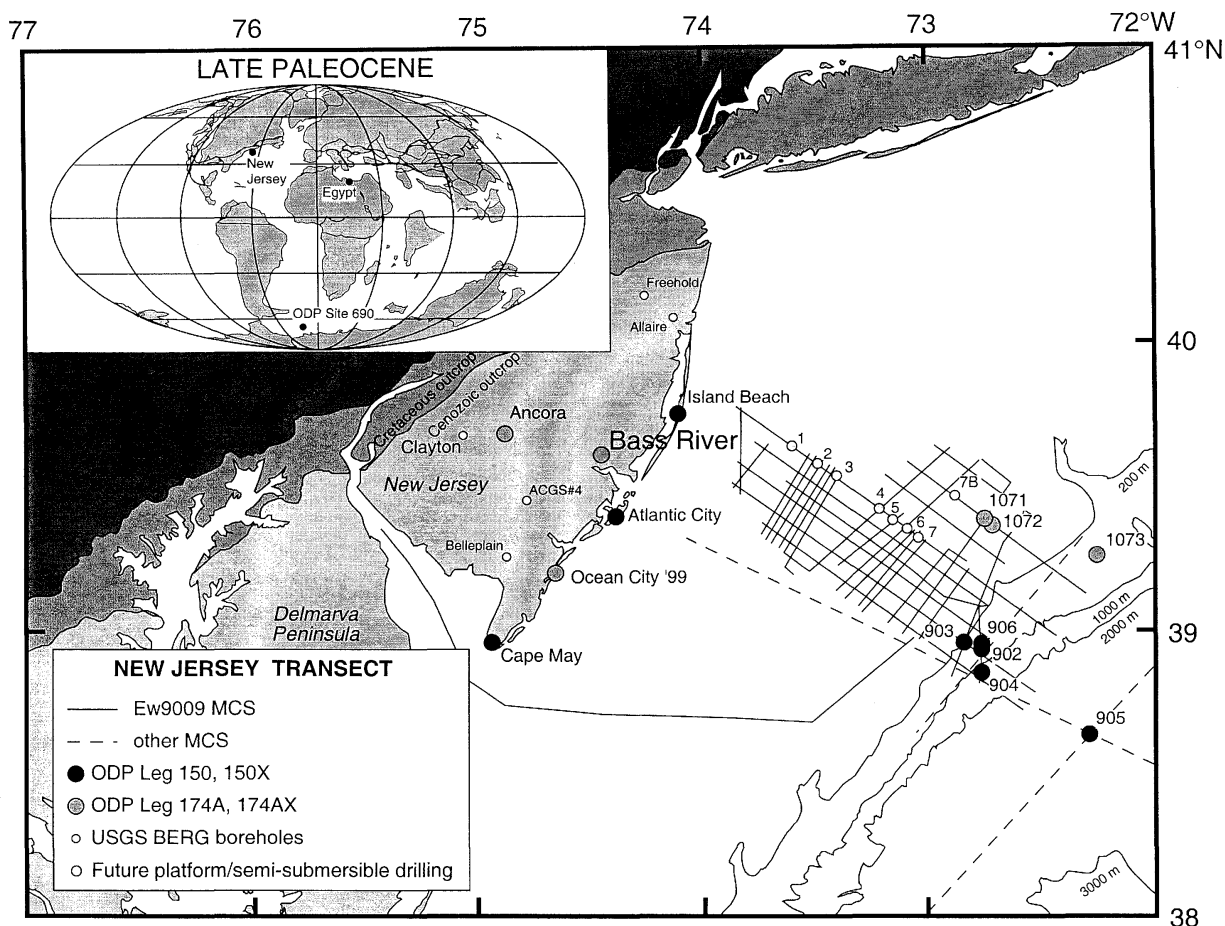


FIG. 1. – Site map, showing location of all onshore and offshore New Jersey boreholes. Inset shows a paleogeographic reconstruction for the late Paleocene [after Thomas, 1998] showing the location of New Jersey, ODP Site 690, and Egypt sections.

FIG. 1. – Carte du New Jersey, donnant la position géographique des sites carottés en mer et à terre. Encadré : reconstitution de la paléogéographie au Paléocène récent [d'après Thomas, 1998] montrant la position du New Jersey, du Site ODP 690 et des coupes égyptiennes.

carbon required to account for the large excursion in isotopic composition of marine dissolved carbonate and atmospheric carbon dioxide has so far only been satisfactorily explained by the dissociation of sediment methane hydrate [Dickens *et al.*, 1995, 1997]. According to this hypothesis, long-term warming of oceanic bottom waters throughout the late Paleocene [Shackleton, 1986; Miller *et al.*, 1987b] led to a decreased sediment thermal gradient which caused an increase in the depth of the methane hydrate stability field. Dissociation of previously stable methane hydrate could then result in a runaway greenhouse effect.

The LPTM is of interest stratigraphically as a candidate for the (re)definition of the Paleocene/Eocene (P/E) boundary and from the perspective of climate research as an example of intense, rapid global warming. Consequently, it has been the subject of a burgeoning literature detailing the stratigraphic and paleoenvironmental records of sections throughout the world. This paper presents benthic foraminiferal carbon and oxygen isotope and clay mineral records from an apparently complete LPTM section recently drilled (October–November, 1996) at Bass River State Forest, New Jersey (fig. 1) as part of the New Jersey Coastal Plain Drill-

ing Project, ODP Leg 174AX [Miller, Sugarman, Browning *et al.*, 1998].

## BACKGROUND

### Chronostratigraphy

The CIE occurs within planktonic foraminiferal Zone P5, calcareous nannoplankton Zone NP9, and magnetic polarity Chron C24r [Berggren and Aubry, 1998]. The P/E boundary is currently defined as the base of the Ypresian Stage (base of the Ieper Clay, Belgium) which is correlative with the calcareous nannoplankton NP10a/b subzonal boundary [Aubry, 1998]. Nevertheless, many authors have prematurely adopted the use of the CIE and associated stratigraphic markers to characterize the P/E boundary. The P/E boundary may be (re)defined at the stratigraphic level of the CIE in recognition of the variety of chemo- and bio-stratigraphic markers that can serve for long distance correlations in both marine and terrestrial records. In this paper we maintain the current definition of the P/E boundary so as to correlate with the NP10a/b subzonal boundary, placing the CIE in the latest Paleocene.

### Climate and Environment

The carbon isotope excursion (CIE) was first described by Kennett and Stott [1990, 1991] at ODP Sites 689 and 690 on Maud Rise in the Southern Ocean. They noted two characteristics of the event: first, both carbon and oxygen isotopes rapidly reached more negative values than have been recorded at any other time during the Cenozoic, and second, the surface to deep water oxygen isotope gradient became reversed during the excursion at the deeper-water Site 690 compared to Site 689. They interpreted the isotope record as indicating a brief warm episode reaching to high latitudes. High latitude warming was coincident with formation of warm saline deep waters at low latitudes, causing a reversal in the oxygen isotope gradient at high latitudes. Although numerous isotope records of the LPTM from various marine sections have subsequently been published [e.g., Pak and Miller, 1992; Lu and Keller, 1993; Bralower *et al.*, 1995; Thomas and Shackleton, 1996], the record from Site 690 remains the most stratigraphically expanded record of the excursion [see Aubry *et al.*, 1996, 1998 for discussion].

The argument that high latitudes experienced "subtropical" warm conditions during the LPTM has been strengthened by Robert and Kennett [1992, 1994], who documented a gradual rise in the proportion of kaolinite in clays deposited through the late Paleocene and early Eocene at Sites 689 and 690. They interpreted this rise as indicative of increased rainfall during the late Paleocene [Robert and Kennett, 1992]. The broad rise is punctuated by a sharp peak in kaolinite/illite and smectite/illite ratios which is coincident with the CIE and was interpreted as indicating abruptly increased precipitation induced by an increased continent-to-ocean temperature gradient [Robert and Kennett, 1994].

### Neritic records

In the neritic environment the CIE has been recorded at Gebel Aweina and Gebel Duwi, Egypt [Schmitz *et al.*, 1996]. Due to severe diagenetic infilling of foraminiferal tests, Schmitz *et al.* [1996] generated benthic foraminiferal records using *Lenticulina* spp. and *Fronidularia* spp., taxa having a thick test which allowed infilling calcite to be removed with the use of a needle and sonication. The record from Egypt shows lower  $\delta^{13}\text{C}$  values than the record from ODP 690 with a less pronounced rebound after the CIE. Schmitz *et al.* [1996] interpreted this as not supporting low latitude formation of bottom waters from this part of the Tethys Ocean, because carbon isotope values would then be expected to decrease toward high latitudes. The oxygen isotope record from Egypt does not show a substantial excursion, which led Schmitz *et al.* [1996] to argue against substantial low-latitude warming and an enhanced greenhouse effect during the CIE. The oxygen isotope record at ODP Site 865 (equatorial Pacific) [Bralower *et al.*, 1995] also suggests that tropical sea surface temperatures remained stable through the LPTM. However, despite the great care taken by Schmitz *et al.* [1996] for sample preparation, we cannot exclude the possibility that the record from Egypt may reflect an incomplete section and/or diagenetic alteration of original isotopic values.

Upper Paleocene-lower Eocene sections were continuously cored at Clayton and Island Beach, New Jersey (fig. 1). Gibson *et al.* [1993] noted several significant changes of paleoenvironmental indicators in the Clayton section at a level stratigraphically correlative with the CIE which has since been shown to contain a  $\sim 8\%$  negative excursion in bulk carbonate carbon and oxygen isotopes [Thomas *et al.*, 1997]. In the Clayton section the lithology changes from a clayey, glauconitic sand (40 + % sand) to a non-glauconitic

clay ( $\sim 1\%$  sand) in the interval containing the CIE. The same level marks the beginning of a rise in kaolinite content in the sediment clay fraction which extends  $\sim 9$  m ( $\sim 30$  ft) above the CIE. Kaolinite makes up 4-5 % of the clay fraction below the CIE and rises to  $\sim 50\%$  by 3 m (10 ft) above the CIE. Gibson *et al.* [1993] interpreted these and other events as indicative of increased precipitation and higher temperatures accompanied by low-oxygen bottom-water conditions and an increase in water depth along the New Jersey coast. Pak *et al.* [1997] generated an isotope record through the upper Paleocene-lower Eocene at Island Beach and correlated a negative excursion in carbon and oxygen isotopes near the P/E boundary with the CIE; however, the exact stratigraphic correlation of the excursion was uncertain because of an unconformity at this level.

### BASS RIVER BOREHOLE

The Bass River core was drilled by Rutgers University and the New Jersey Geological Survey during October-November, 1996, and is the first site drilled as part of ODP Leg 174AX, complementing shelf drilling by Leg 174A [Austin, Christie-Blick, Malone *et al.*, 1998]. The site was the fourth hole drilled as part of the New Jersey Coastal Plain Drilling Project, following Leg 150X drilling at Island Beach, Cape May, and Atlantic City [Miller *et al.*, 1994a, b, 1998] (fig. 1). A full description of the Bass River core and onsite operations is published in Miller, Sugarman, Browning *et al.* [1998]. Paleomagnetic (both AF demagnetization and Kirschvink analysis), clay mineral, stable isotope, and sediment data generated for this study are available from the World Data Center-A for Paleoclimatology, 325 Broadway, Boulder, CO, USA; <http://www.ngdc.noaa.gov/paleo/paleo.html>; email: [paleo@ngdc.noaa.gov](mailto:paleo@ngdc.noaa.gov).

### Lithostratigraphy

The lithostratigraphy of the upper Paleocene-lowermost Eocene interval in New Jersey has been somewhat uncertain. Olsson and Wise [1987a, b] noted the presence of a "gray clay, silt, and sand unit" in the subsurface separating the upper Paleocene Vincentown Formation from the lower Eocene Manasquan Formation. Gibson *et al.* [1993] correlated this unit with a section in the Clayton core which they assigned to the Manasquan Formation. They place the Vincentown/Manasquan contact at the lithologic change where the CIE has been subsequently shown to occur [Thomas *et al.*, 1997]. We note that this unit at Clayton is not typical Manasquan lithology (glauconitic sandy green clay) and that the interval below the CIE is not typical Vincentown lithology (quartz sand to glauconite sand). Liu *et al.* [1997] correlated an upper Paleocene unit of interbedded gray clay and sandy silt in the Island Beach core with the unnamed unit of Olsson and Wise [1987a, b]. Owens *et al.* [1997] assigned this unit to the Vincentown Formation. We adopt the terminology of Olsson and Wise [1987a, b], assigning gray clays, silts, and fine sands of the uppermost Paleocene in New Jersey to an unnamed unit lithologically distinct from the Vincentown Formation below and the Manasquan Formation above.

Miller, Sugarman, Browning *et al.* [1998] locate two depositional sequences within the upper Paleocene Vincentown Formation, in which they include the unnamed unit of Olsson and Wise [1987a, b]: 1138.6-1240.8 ft (347.05-378.20 m) and 1243.9-1256.5 ft (379.14-382.98 m) (fig. 2). Surfaces were observed in the core at 1240.8 and 1243.9 ft (378.20 and 379.14 m), either of which could be the boundary between these two sequences. Sediment becomes glau-

conitic and sandy below ~ 1236 ft (~ 376.7 m), just above the sequence boundary, and we assign this interval and the glauconitic muds between 1241.5 and 1248.5 ft (378.41 and 380.54 m) to the Vincentown Formation. Above the basal glauconitic sandy silts the uppermost Paleocene sequence (1138.6-1236 ft, 356.92-376.73 m) is a very slightly sandy, slightly glauconitic silty gray clay containing evidence of bioturbation and very few sedimentary structures which becomes a slightly silty greenish-gray clay above ~ 1171 ft (~ 356.9 m). We correlate these clays and silty clays with the unnamed unit of Olsson and Wise [1987a, b] and Liu *et al.* [1993], (Island Beach borehole) and the upper Vincentown-Manasquan Formations of Gibson *et al.* [1993, Clayton borehole] (fig. 2). A distinct unconformity at 1138.6 ft (347.05 m) separates the uppermost Paleocene unnamed unit from the lower Eocene Manasquan Formation. A very thin (4.0 ft, 1.22 m) sequence occurs in the lowermost Manasquan with an upper sequence boundary marked by a surface at 1134.6 ft (345.83 m).

### Core sampling

Samples for planktonic foraminiferal, isotope, and clay mineral analysis, taken at a sampling interval of 1-5 ft. (30-150 cm), were dried in a 40 °C oven and weighed dry. Samples were disaggregated in sodium metaphosphate at room temperature and washed through a 63 µm screen. The fine (< 63 µm) fraction was saved for use in preparing samples for clay analysis. The sand (> 63 µm) fraction was dried in a 40 °C oven and weighted to determine sediment weight percent sand. Splits were separated from the sand fraction for isotope analysis. Samples for calcareous nanofossil studies were taken at a sampling interval of ~ 1 ft (~ 30 cm), with increased sample density in the vicinity of distinct lithologic surfaces.

Ten to fifteen cm<sup>3</sup> quarter-round sections of core were taken for magnetostratigraphic analysis. Samples were taken at 5-10 ft (150-300 cm) intervals through the section with additional samples at 1-2 ft (30-60 cm) intervals to better resolve magnetozone boundaries.

### Biostratigraphy

Calcareous nannoplankton biostratigraphy (fig. 2) is discussed in Miller, Sugarman, Browning *et al.* [1998] and Aubry *et al.* (manuscript in prep.). Preservation of calcareous nanofossils is excellent throughout the upper Paleocene-lower Eocene. The interval between 1242.5 and 1244 ft (378.71 and 379.13 m) belongs to Zone NP6 while the interval between 1211.4 and 1240.0 ft (370.36 and 377.95 m) belongs to Zone NP8. Zone NP7 is missing, suggesting that the surface noted at 1240.8 ft (378.20 m) is an unconformity. The base of Zone NP9 is placed between samples at 1210.6 and 1211.4 ft (368.99 and 369.24 m) based on the lowest occurrence (LO) of *Discoaster multiradiatus* at 1210.6 ft (368.99 m). Based on the highest occurrence (HO) of *Helicolithus megastypus* at 1171.1 ft (356.95 m) and the LO of *Rhombaster calcitrata*, *Discoaster araneus*, and *D. anartios* at 1170.9 ft (356.89 m), Aubry *et al.* (manuscript in prep.) subdivide Zone NP9 into Subzones NP9a and NP9b. Zone NP9 is truncated by the unconformity at 1138.6 ft (347.05 m). The juxtaposition of the LOs of *Tribrachiatius digitalis* and *T. bramlettei* at 1138.4 ft (346.98 m) indicates that the sediment immediately above the unconformity belongs to Subzone NP10b [subzonal scheme of Aubry, 1996]. Subzone NP10b is restricted to a very thin interval, with the LO of *T. contortus* at 1138.1 ft (346.89 m) indicating Subzone NP10d. Subzone NP10c is missing, suggesting that there is a hiatus which may cor-

respond to a subtle surface in the core noted at 1138.25 ft (346.94 m). The HO of *T. contortus* at 1136.2 ft (346.31 m) determines the top of Subzone NP10d. The NP10/NP11 contact is unconformable.

Planktonic foraminiferal biostratigraphy (fig. 2) is discussed in Miller, Sugarman, Browning *et al.* [1998]. Planktonic foraminiferal zonal boundaries are uncertain due to poor preservation of foraminifera between samples at 1225.3 and 1170.0 ft (373.47 and 356.62 m) and inconsistencies in the occurrence of datums between 1230 and 1245.4 ft (374.90 and 379.60 m). Planktonic foraminiferal Zone P4 is identified in samples from 1225.3 to 1247.5 ft (373.47 to 380.24 m). The base of Subzone P4c is identified at 1230.0 ft (374.90 m) by the LO of *Acarinina soldadoensis* and the top of Subzone P4a is identified at 1247.5 ft (380.24 m) by the HO of *Acarinina subsphaerica*. The interval from 1234.3 to 1245.4 ft (376.22 to 379.60 m) is assigned to Subzone P4b, although the LO of *M. aequa*, generally found in association with the LO of *A. soldadoensis*, at 1245.4 ft (379.60 m) suggests that this interval may actually belong to Subzone P4c. In either case, this conflicts with identification of calcareous nanofossil Zone NP6 below 1242.5 ft (378.71 m). The HO of *G. pseudomenardi* at 1225.3 ft (373.47 m) marks the top of Zone P4. Poor preservation of planktonic foraminifera between samples at 1225.3 and 1170.0 ft (373.47 and 356.62 m) prevents confident assignment of this interval to Zone P5. The interval from 1145 to 1170 ft (349.00-356.62 m) is confidently assigned to Zone P5 based on the presence of *Morozovella aequa*, *M. gracilis*, *M. subbotinae*, *M. velascoensis*, and *Acarinina soldadoensis*. In addition, three species (*Acarinina africana*, *A. sibiyaensis*, and *Morozovella allisonensis*) that are associated with the LPTM at ODP Site 865 in the tropical Pacific [Kelly *et al.*, 1996] occur in the interval beginning at 1171 ft (356.9 m). *Acarinina africana* and *A. sibiyaensis* are found throughout the interval from 1171 ft to 1145 ft (356.9 to 349.0 m) whereas *M. allisonensis* is confined to a thin interval from 1171 to 1165 ft (356.9 to 355.1 m). Sediments between 1110 and 1130 ft (338.33 and 344.42 m) are assigned to Zone P6b based on the LO of *Morozovella formosa formosa* at 1130 ft (344.42 m). A sample at 1137 ft (346.56 m) is assigned to lowermost Eocene Zone P6a, based on the absence of *Morozovella velascoensis* and *M. formosa formosa* and the co-occurrence of *Pseudohastigerina wilcoxensis*, *Morozovella formosa gracilis*, and *Morozovella aequa*.

### Magnetostratigraphy

Natural remanent magnetism (NRM) was measured on a 2G DC SQUID 3-axis cryogenic magnetometer at Lamont-Doherty Earth Observatory. Step demagnetization of NRM at 5 mT increments to 45 mT was performed using a large-volume alternating field (AF) demagnetization coil. The most stable paleomagnetic direction was calculated from linear demagnetization trends using the method of least-squares analysis [Kirschvink, 1980]. In most cases a series of 3-4 demagnetization steps between 20 and 45 mT were used including the origin; demagnetization patterns indicate that any overprints are removed by around 20 mT, allowing the characteristic magnetization to be isolated.

Paleomagnetic data were divided into three quality categories based on the demagnetization pattern (fig. 2). (1) Good demagnetization patterns with linear demagnetization trends (MAD angles < 5°) are typical of samples from most of the clay-rich uppermost Paleocene sequence. (2) Noisy demagnetization trajectories which nonetheless allow at least the polarity of the characteristic component to be reasonably determined (MAD angles from 5° to 15°)

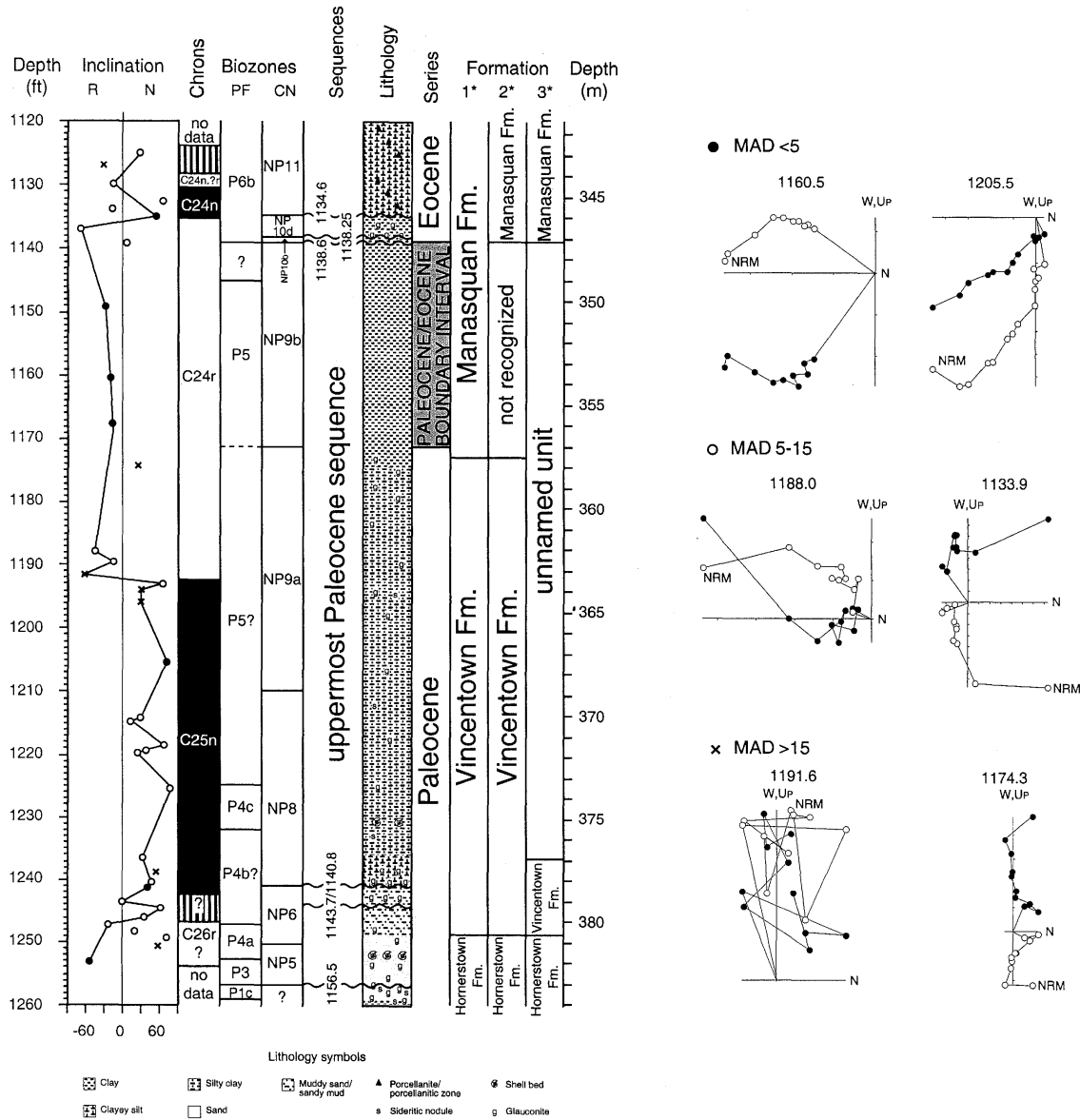


FIG. 2. - Lithologic column showing paleomagnetic data with interpreted chrono boundaries and planktonic foraminiferal [PF, Olsson in Miller, Sugarman, Browning *et al.*, 1998] and calcareous nannoplankton [CN, Aubry in Miller, Sugarman, Browning *et al.*, 1998 and Aubry *et al.*, manuscript in prep.] zonations. Examples of demagnetization patterns are provided, showing the typical quality of data. Demagnetization plots are plotted on an orthogonal axis, with the horizontal axis representing "North" and the vertical axis representing "Up" for open circles and "West" for closed circles. Cores were not oriented in the horizontal plane, so declination data (closed circles) are inconsistent between samples. Three formational nomenclatures are shown: 1\* after assignment of Gibson *et al.* [1993] at Clayton; 2\* after assignment of Owens *et al.* [1997] at Island Beach; 3\* this study, after Olsson and Wise [1987a, b] and after assignment of Liu *et al.* [1997] at Island Beach.

FIG. 2. - Corrélations lithologiques, magnéto- et biostratigraphiques dans la coupe de Bass River. biozonations à partir des foraminifères planctoniques (PF) : Olsson in Miller, Sugarman, Browning *et al.* [1998]; à partir des nannofossiles calcaires (CN) : Aubry in Miller, Sugarman, Browning *et al.* [1998] and Aubry *et al.*, soumis. Deux exemples de démagnétisation progressive typique sont donnés pour rendre compte de la qualité des données. Les valeurs de démagnétisation sont reportées sur un diagramme, dont l'axe horizontal représente le Nord et l'axe vertical représente « up » pour les cercles ouverts et « ouest » pour les cercles fermés. Les carottes n'étant pas orientées par rapport à l'horizontale, les valeurs de la déclinaison (cercles fermés) sont très variables. Trois interprétations lithostratigraphiques sont données : 1\* d'après Gibson *et al.* [1993] basée sur Clayton; 2\* d'après Owens *et al.* [1997] basée sur Island Beach; 3\* cette étude, d'après Olsson et Wise [1987a, b] et d'après Liu *et al.* [1997] basée sur Island Beach.

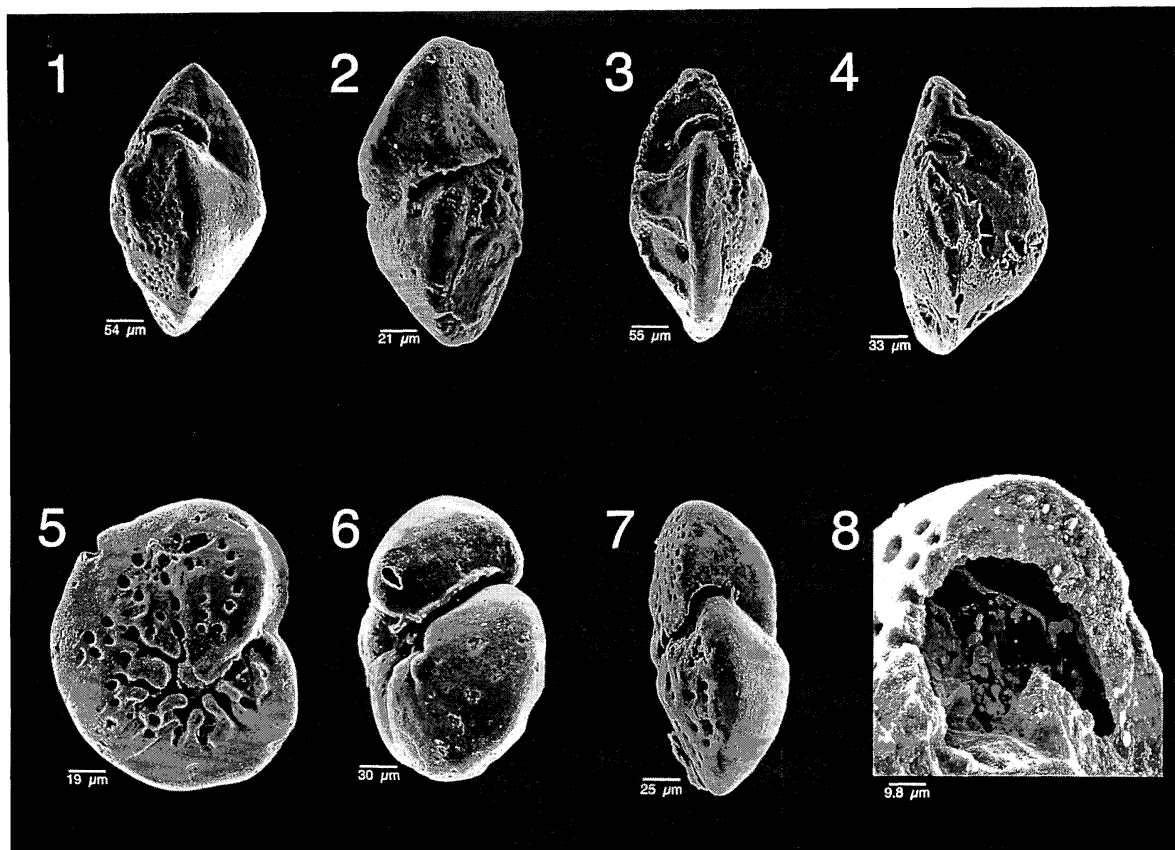


FIG. 3. – Scanning electron micrographs showing preservation of benthic foraminifera. 1. *C. alleni* (1180.0 ft, 359.66 m); 2. *C. alleni* (1163.0 ft, 354.48 m); 3. *C. succedens* (1161.0 ft, 353.87 m); 4. *C. succedens* (1183.0 ft, 360.58 m); 5. *S. beccariiformis* (1183.0 ft, 360.58 m); 6. *G. octocamerata* (1165.0 ft, 355.09 m); 7. *A. acuta* (1170.4 ft, 356.74 m); 8. close-up of broken *A. acuta* test (1165.0 ft, 355.09 m). Note the difference in preservation of specimens from above and below 1171.0 ft (356.92 m).

FIG. 3. – Microphotographies (MEB) montrant l'état de conservation des tests des foraminifères benthiques. 1. *C. alleni* (1180,0 ft, 359,66 m); 2. *C. alleni* (1163,0 ft, 354,48 m); 3. *C. succedens* (1161,0 ft, 353,87 m); 4. *C. succedens* (1183,0 ft, 360,58 m); 5. *S. beccariiformis* (1183,0 ft, 360,58 m); 6. *G. octocamerata* (1165,0 ft, 355,09 m); 7. *A. acuta* (1170,4 ft, 356,74 m); 8. détail d'un test cassé de *A. acuta* (1165,0 ft, 355,09 m). Notez la différence d'état de conservation des tests au dessus et en dessous de 1171,0 ft (356,92 m).

are typical of samples from sandier portions of the section. (3) Demagnetization patterns showing hardly any linearity (MAD angles  $> 15^\circ$ ) often appear to be associated with intervals surrounding a reversal in magnetic polarity.

Based on clearly reversed samples at 1247.2 and 1253.1 ft (380.15 and 381.24 m) the interval from 1247.2 to 1253.1 ft (380.15 to 381.24 m) is tentatively interpreted as a reversed polarity interval (fig. 2). Integration with biostratigraphy suggests that this interval represents Chron C26r. Samples between 1243.7 and 1246.0 ft (380.15 and 379.78 m) may be overprinted and are not interpreted. All samples from 1193.00-1241.30 ft (363.63-378.35 m) show normal polarity and this interval is confidently interpreted as representing Chron C25n, although we cannot completely rule out overprinting. We note that this conflicts with assignment of the interval from 1234.3 to 1241.3 ft (376.22 to 378.35 m) to planktonic foraminiferal Zone P4b. Samples between 1137.00 and 1191.60 ft (346.56 and 363.20 m) show reversed polarity except for a sample at 1174.30 ft (357.93 m) which shows evidence of overprinting. The reversed polarity interval from 1137.0 to 1191.6 ft (346.56 to 363.20 m) is confidently interpreted as rep-

resenting Chron C24r. Samples above 1135.00 ft (345.19 and 345.95 m) show inconsistent polarity and are generally very weakly magnetized. We confidently interpret samples at 1125.0 and 1135.0 ft (342.90 and 345.95 m) as normal polarity and one sample at 1129.9 ft (344.39 m) as reversed polarity and tentatively correlate this interval with Chron C24n with possibly either C24n.1r or C24n.2r represented at 1129.9 ft (344.39 m).

#### Isotope analysis

Samples of *Cibicidoides* spp. (mostly *C. succedens*, *C. alleni* and *C. cf. vulgaris*), *Anomalinoidea acuta*, *Gyroidinoidea octocamerata* and *Stensioina (Gavelinella) beccariiformis* were picked for carbon and oxygen isotope analysis. Sample spacing was 5-10 ft (150-300 cm) below 1190 ft (362.7 m) and above 1160 ft (353.6 m) and 1-2 ft (30-60 cm) between 1160 and 1190 ft (353.6 and 362.7 m). Above 1173 ft (357.53 m) preservation is excellent; below 1173 ft (357.53 m) preservation varies from poor to very poor (fig. 3). Very poorly preserved specimens of *C. succedens* lacking an aperture are virtually indistinguishable

from poorly preserved *Nuttalides truempyi*. Although every effort was made to run only specimens which could be confidently identified, low abundances required that some specimens were run which lacked an aperture. It is not certain that these samples were strictly *Cibicidoides* spp. and this may account for the variability in the isotope record (especially oxygen) below 1173 ft (357 m). Published values for offsets of *Nuttalides* from *Cibicidoides* are 0.0 ‰  $\delta^{13}\text{C}$  and -0.15 ‰  $\delta^{18}\text{O}$  [Shackleton *et al.*, 1984] and -0.12 ‰  $\delta^{13}\text{C}$  and -0.23 ‰  $\delta^{18}\text{O}$  [Pak and Miller, 1992], so any mixture of genera should have a minimal effect on the isotope records.

Most samples of *Cibicidoides* spp., *A. acuta*, and *G. octocamerata* were sonicated for <2 sec in distilled water. *Stensioina beccariiformis* samples and some *Cibicidoides* spp. samples below 1173 ft (357 m; just below the base of the CIE) were not sonicated due to the fragility of the specimens. Isotope analysis was performed at the University of Maine, Orono. The samples were reacted in phosphoric acid ( $\text{H}_3\text{PO}_4$ ) at 90 °C in an Autocarb peripheral attached to a Micromass Prism II mass spectrometer. The analytical precision of the NBS19 and NBS20 standards analyzed during the sample runs was 0.06 and 0.05 ‰ for  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  values, respectively.

The foraminiferal isotopic record (fig. 4) is a composite of spliced, overlapping records using assemblages of *Cibicidoides* spp., *A. acuta*, *G. octocamerata*, and *S. beccariiformis*. Values for *Cibicidoides* spp. are ~ -1 ‰ in oxygen and ~ 1 ‰ in carbon below 1173 ft (357.53 m). A major negative excursion in both carbon and oxygen isotopes occurs beginning at 1173.0 ft (357.53 m) with the lowest values (-3.18 ‰  $\delta^{13}\text{C}$ , -3.69 ‰  $\delta^{18}\text{O}$ ) occurring at 1171.1 ft (356.95 m). Between 1173.0-1171.1 ft (357.53-356.95 m) the magnitude of the excursion in *Cibicidoides* spp. is -4.0 ‰ in  $\delta^{13}\text{C}$  and -2.3 ‰ in  $\delta^{18}\text{O}$ . Between 1169 and 1146 ft (356.31 and 349.3 m) values recover gradually from -3 to -1 ‰ in carbon and from -3 to -2 ‰ in oxygen.

Splicing of the isotopic record is necessary because of an abrupt change in the benthic assemblage between 1173.0 and 1171.1 ft (357.53 and 356.95 m) [Cramer, 1998 and unpublished]. No specimens of *S. beccariiformis* are present above 1171.1 ft (356.95 m) and no specimens of *G. octocamerata* or *A. acuta* are present below 1171.8 ft (357.16 m). There is a correlative change in *Cibicidoides* species assemblage. *Cibicidoides* spp. samples below 1173.0 and above 1165.0 ft (below 357.53, above 355.09 m) are generally a mixture of *C. allenii* and *C. succedens* while those between 1171.8 and 1166.0 ft (357.16-355.40 m) are generally of *C. cf. vulgaris*. We note in this context that the relatively high values in both carbon and oxygen isotopes (-0.85 ‰ and -2.18 ‰, respectively) in one *Cibicidoides* spp. sample at 1171.1 ft (356.95 m) was obtained using a monospecific sample of *C. succedens*. *Cibicidoides succedens* are very rare in the interval from 1171.1 to 1165.0 ft (356.95 to 355.09 m) and it is likely that all *C. succedens* specimens found in this interval were bioturbated upward from below the level of the excursion and therefore show relatively high isotopic values.

#### Clay mineral analysis

Clay analysis was run on the ~ <2  $\mu\text{m}$  fraction at a sampling interval of 1-4 ft (30-120 cm) from 1136.2-1219.4 ft (346.31-371.67 m). The < 63  $\mu\text{m}$  sieved fraction was placed in 1 L settling columns filled with distilled water. The < 2  $\mu\text{m}$  size fraction was obtained by allowing the suspension to settle for 3.5 hours at room temperature and pipetting off the upper 4 cm of liquid. The clay fraction was treated with 5 % acetic acid to remove carbonate and allowed to

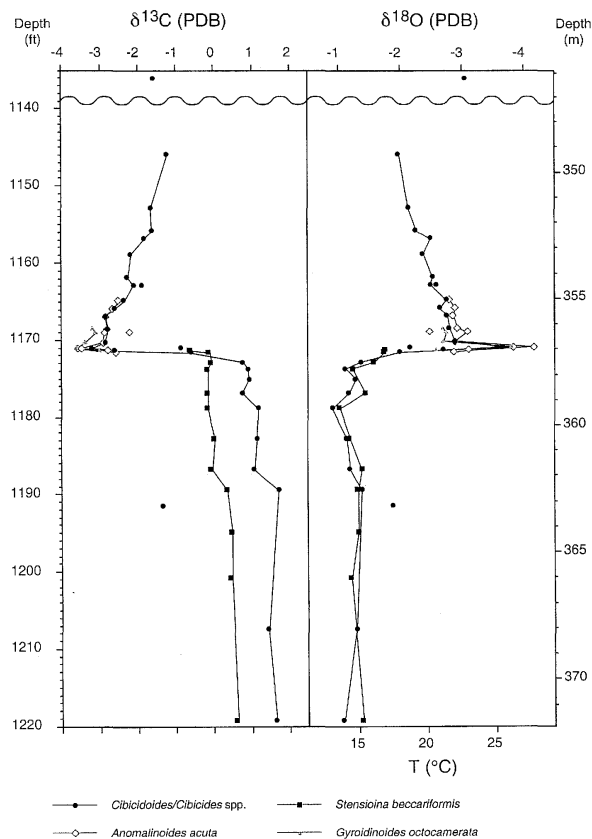


FIG. 4. — Carbon and oxygen isotope records from Bass River. Temperature is calculated for *Cibicidoides* spp. based on the equation  $T = 16.9 - 4.38(\delta_c - \delta_w) + 0.10(\delta_c - \delta_w)^2$  [O'Neil *et al.*, 1969; Shackleton, 1974] assuming an ice free world with  $\delta_w = -1.2$  ‰ [Shackleton and Kennett, 1975] and that *Cibicidoides* spp. secretes calcite which is depleted by 0.64 ‰ relative to  $\delta^{18}\text{O}$  equilibrium [Graham *et al.*, 1981].

FIG. 4. — Isotopes du carbone et de l'oxygène dans la coupe de Bass River. La température déterminée sur la base de *Cibicidoides* spp. est établie à partir de l'équation  $T = 16.9 - 4.38(\delta_c - \delta_w) + 0.10(\delta_c - \delta_w)^2$  [O'Neil *et al.*, 1969; Shackleton, 1974] en admettant qu'il n'y avait de glace, avec  $\delta_w = -1.2$  ‰ [Shackleton et Kennett, 1975] et que *Cibicidoides* spp. sécrète une calcite appauvrie de 0.64 ‰ par rapport à l'équilibre du  $\delta^{18}\text{O}$  [Graham *et al.*, 1981].

dry to a paste. Smear slides were prepared from the paste and the surface of the slide was smoothed with a spatula. This method has been found to achieve well oriented samples which yield diffraction patterns nearly identical to those obtained by suction onto a ceramic tile [Theisen and Harward, 1962; Gibbs, 1965]. The slides were analyzed dry and after glycol solvation at the New Jersey Geological Survey on a Diano 2100e X-ray diffractometer (XRD) using a Cu target. Counts per second (CPS) were measured at 0.04 °2 $\theta$  increments from 5-33 °2 $\theta$ .

The clay assemblage consists primarily of kaolinite and a mixed layer illite/smectite (> 80 % illite), possibly with a minor component of illite. Illite and the mixed-layer illite/smectite are treated as a single illite/smectite component. It is possible that there is a glauconite component present, which shares some refraction angles with illite. A glauconite 001 refraction should occur at the same angle

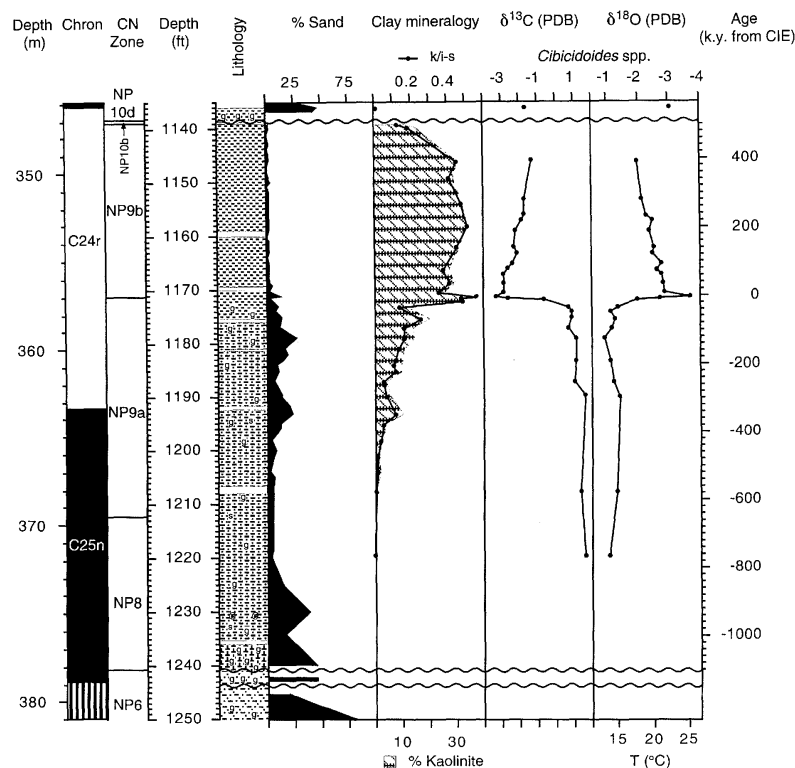


FIG. 5. – Environmental indicators through the Upper Paleocene sequence at Bass River. Temperature was calculated as in figure 4. Age was calculated based on a sedimentation rate of 19.3 m/m.y. See Figure 2 for lithology symbols.

FIG. 5. – Indicateurs paléoenvironnementaux dans la séquence d'âge paléocène supérieur de Bass River. Température calculée comme pour la figure 4. Ages établis sur la base d'un taux de sédimentation de 19.3 m/m.y. Voir la figure 2 concernant la signification des symboles lithologiques.

as illite 001; however, no significant glauconite refraction peak should occur at the same angle as illite 002 [Moore and Reynolds, 1989]. Therefore, any significant variation through the section in the amount of glauconite clay present should cause variation in the ratio of the intensities of the illite 001 and 002 peaks (which are expressed as illite-smectite 001/004\* and 002/003 peaks in this section). No discernible variation is observed in this ratio, so we assume that glauconite does not contribute significantly to variation in clay mineral abundances.

Quantitative analysis was performed on the glycol saturated traces using procedures described in Moore and Reynolds [1989]. All ratios of kaolinite to illite/smectite peaks (kaolinite 001 and 002 and illite/smectite 001/004\*, 002/003 and 003/005) show the same variation through the section, indicating that the results of the quantitative analysis are robust. For details of clay mineral analysis procedures see Cramer [1998].

At 1214.5 ft (370.18 m), kaolinite is essentially absent from the clay assemblage (fig. 5). Kaolinite content gradually rises to ~ 34 % at 1158.0 ft (352.96 m) and declines to < 2 % at the upper unconformity (1138.7 ft, 347.08 m). Superimposed on the gradual rise is a sharp spike between 1171.8 ft and 1171.1 ft (357.16-356.95 m) with values of 32-36 % kaolinite (figs. 5, 6). This spike indicates a pulse of kaolinite coincident with the CIE which increased the

kaolinite content of the sediment being deposited at this site by > 10 % over the first-order variation.

### Coarse fraction

The percentage of sand in the sediment was calculated based on the weight of samples before and after washing through a 63  $\mu\text{m}$  sieve. Relative proportions of components of the sand fraction were determined by visual examination under a light microscope. The total coarse fraction (sand-sized, > 63  $\mu\text{m}$ ) below 1173.0 ft (357.53 m) varies between 5-20 % (fig. 5). From 1171.8 ft (357.16 m) to the upper unconformity (1138.7 ft, 347.08 m), the total sand fraction is only ~ 1 %, occasionally increasing to 3 %. The one exception is the sample at 1171.1 ft (356.95 m) which contains ~ 10.3 % sand-sized material, nearly all of which is (predominantly planktonic) foraminifera (fig. 6). From 1171.1 ft (356.95 m) to the top of the section, foraminifera make up 80-100 % of the sand fraction, while below 1175.2 ft (358.2) the sand fraction is dominantly glauconite and quartz. Glauconite grains in this interval are well sorted and of the same size and shape as quartz grains, indicating that the glauconite and quartz were transported by the same mechanism. Glauconite is therefore treated as a terrigenous component derived from older exposed marine deposits rather than an authigenic component produced during the late Paleocene.



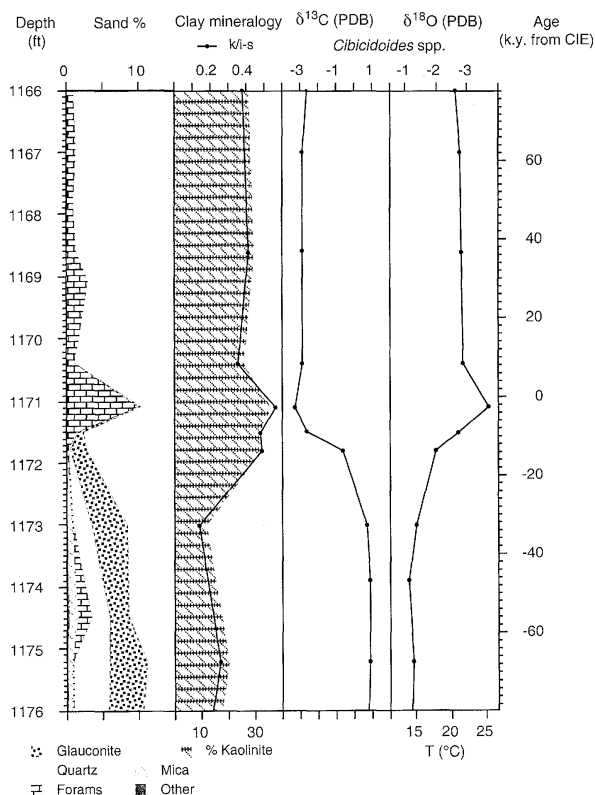


FIG. 6. – Environmental indicators across the latest Paleocene thermal maximum at Bass River. See figure 5 for explanation.

FIG. 6. – Indicateurs paléoenvironnementaux dans la séquence d'âge paléocène supérieur de Bass River. Voir figure 5 pour explication.

## DISCUSSION

### Stratigraphy

Integration of paleomagnetic [Cramer, 1998; this study], calcareous nannofossil [Aubry in Miller, Sugarman, Browning *et al.*, 1998 and Aubry *et al.*, manuscript in prep.], and planktonic foraminiferal [Olsson, in Miller, Sugarman, Browning *et al.*, 1998] zonations provide a reliable chronology for the uppermost Paleocene sequence at Bass River (figs. 2, 7). In constructing an age model, we use the Chron C25n/C24r reversal, the LO of *D. multiradiatus* (marking the base of nannofossil Zone NP9), and the CIE, which essentially coincides with the LO of *R. calcitrapa* (base of nannofossil Subzone NP9b) (table I, fig. 7). Our rationale in choosing these three datums is twofold: (1) to pick datums which are most likely to be reliable in the Bass River borehole and which do not directly contradict other stratigraphic datums; and (2) to pick datums which will allow correlation of the Bass River record with the global record of the LPTM. The reversal in magnetic polarity between 1191.6 and 1193.0 ft (363.20 m and 363.63 m) is very clearly delineated (fig. 2) and its occurrence within Zone NP9 indicates that this is the Chron C25n/C24r reversal. The LO of *D. multiradiatus* is clearly delineated in an interval of good nannofossil preservation. The CIE itself (1171.1 ft, 356.95 m) is also used as a chronologic datum, taking the age of 55.52 Ma determined by Aubry *et al.*

TABLE I. – Age model for the latest Paleocene, Bass River.

TABLEAU I. – Interprétation par rapport au temps du Paléocène supérieur, Bass River.

Datums	Depth interval (ft)		Mean depth		Age (Ma)
	lower	upper	(ft)	(m)	
CIE	1171.5	1170.4	1170.95	356.906	55.52
C25n/C24r	1193.0	1191.6	1192.30	363.413	55.904
NP8/NP9	1210.6	1211.4	1211.00	369.113	56.2

[1996] based on the ODP Site 690 stratigraphy relative to the Berggren *et al.* [1995] timescale. The fact that the CIE in the Bass River section falls on a sedimentation rate line extrapolated from the LO of *D. multiradiatus* and the Chron C25n/C24r reversal (fig. 7) lends credence to the age calculated by Aubry *et al.* [1996] and to the use of the CIE as a stratigraphic datum.

In using the Chron C25n/C24r reversal, the LO of *D. multiradiatus*, and the CIE as stratigraphic datums we recognize that these are not entirely consistent with planktonic foraminiferal bio-stratigraphy and magnetostratigraphic interpretation in the interval from ~ 1220 ft (371.86 m) to the unconformity at 1240.8 ft (378.20 m) (fig. 7). The low position of the HO of *G. pseudomenardii* and the high position of the LO of *M. subbotinae* are easily explained by poor preservation of planktonic foraminifera between 1170 and 1225, as noted above. We note that, although an age for the HO of *A. mckannai* is provided in BKSA95, this datum is generally less reliable than the others provided. The LO of *A. soldadoensis* is essentially consistent with a sedimentation rate line extrapolated from the Chron C25n/C24r reversal and the LO of *D. multiradiatus*. However, we hesitate to interpret this LO as representing the FAD of the species due to the LO of *M. aequa* at 1245.4 ft (379.60 m), which should occur in association with the LO of *A. soldadoensis*. In addition, magnetostratigraphic interpretation of Chron C25n extending downward at least to the unconformity at 1240.8 ft (378.20 m) is more consistent with the LO of *M. aequa* at 1245.4 ft (379.60 m) than with the LO of *A. soldadoensis* at 1230.0 ft (374.90 m). However, we cannot rule out the possibility that the magnetic polarity record may be overprinted in this interval. These conflicts may be resolved in future more detailed planktonic foraminiferal and magnetostratigraphic studies.

Our age model indicates a sedimentation rate of ~ 19.3 m/m.y. (fig. 7). The sedimentation rates above the CIE and below the LO of *D. multiradiatus* are poorly constrained. However, for convenience and in order to estimate rates of environmental changes we apply a uniform sedimentation rate of 19.3 m/m.y. throughout the sequence, noting that in reality this rate may vary significantly. This sedimentation rate indicates an age at the upper sequence boundary (1138.7 ft, 347.08 m) of 55.0 Ma which we note is a minimum age since the NP9/NP10 biochronal boundary has an age estimate of 55.0 Ma (BKSA95). We estimate a maximum age for this sequence boundary (1138.7 ft, 347.08 m) of 55.3 Ma based on assuming a reasonable sedimentation rate (< 50 m/m.y.) above the CIE. Extrapolation from the LO of *D. multiradiatus* to the lower sequence boundary (1240.8 ft, 378.20 m) indicates an age of 56.6 Ma. We estimate an age for this sequence boundary (1240.8 ft, 378.20 m) of 56.3-56.6 Ma, with the minimum age estimated assuming that the magnetostratigraphic interpretation is correct. Our age model and these constraints on the ages of the two sequence boundaries allows correlation of the Bass River record of the LPTM with the global record with a high degree of precision.

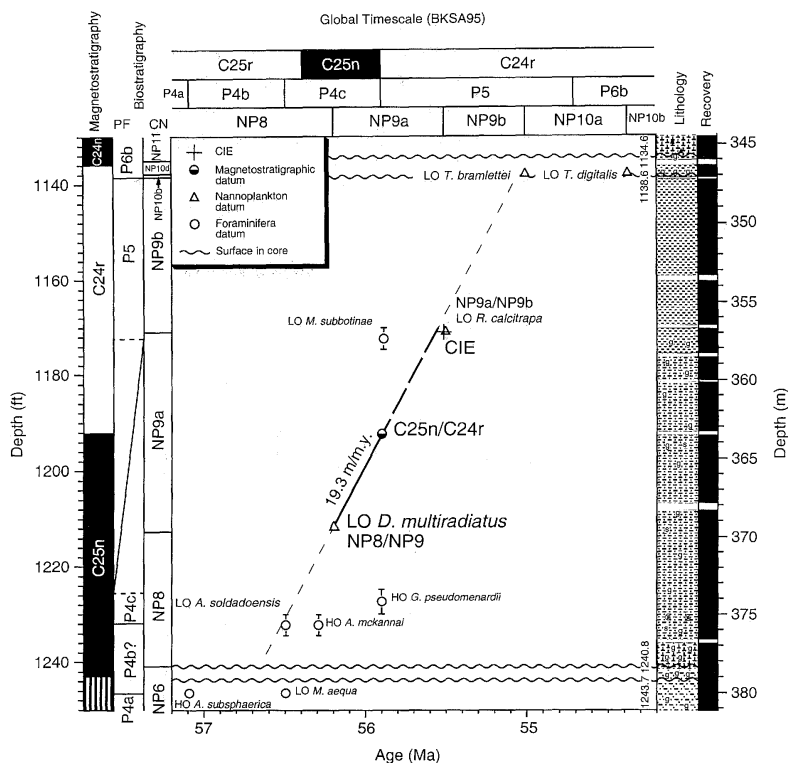


FIG. 7. – Age model for the Upper Paleocene sequence at Bass River. Error bars on plotted points reflect sampling interval. Sedimentation rate of 19.3 m/m.y. was calculated based on the C25n/C24r Chron boundary and the NP8/NP9 Biochron boundary using ages from BKSA95. See text for discussion. Lithology symbols are as in figure 2.

FIG. 7. – Interprétation par rapport au temps de la séquence paléocène de Bass River. Les barres d'erreur reflètent les incertitudes liées au pas d'échantillonnage. Un taux de sédimentation de 19,3 m/m.y. est estimé sur la base de l'inversion magnétique (Chron C24n/C24r) et de la limite biochronale NP9/NP10. Voir le texte pour discussion. Symboles lithologiques comme pour la figure 2.

We recognize, though, that the accuracy of these ages depends on a chronology derived from a composite section for Chron C24r derived using DSDP Site 550 (late Chron C24r) and ODP Site 690 (early Chron C24r) [Aubry *et al.*, 1996]. As recognized by Aubry *et al.* [1996] and Berggren and Aubry [1998], a numerical chronology will be firmly established only when a continuous section allowing accurate estimation of sedimentation rates through Chron C24r is recovered. Moreover, there is currently no means to determine the completeness of sections at or close to the NP9a/b subzonal boundary. The Bass River borehole provides a record of Subzone NP9b and the LPTM, which are preserved in very few other sections [Aubry, 1998]. Although this record is consistent with the Site 550/690 composite section it does not contain a complete section through Chron C24r and we are currently unable to determine if a gap occurs within Subzone NP9a. Yet the correlation that we observe between the isotopic shifts, the kaolinite enrichment, the occurrence of the planktonic foraminifera excursion taxa and that of the LPTM calcareous nannoplankton taxa (Aubry, unpublished data) as seen in several deep sea records, indicates that the Bass River section yields an essentially complete record of the LPTM. Although the relative chronology is firm, we emphasize that the ages and durations calculated based on our age model must be treated with caution until an independent chronology is derived.

### Significance of the Bass River section

The Bass River section provides an apparently complete record of events surrounding the LPTM extending from near the beginning of magnetic polarity Chron C25n to near the end of Biochron NP9, an estimated 800-1100 k.y. prior to and 200-500 k.y. after the CIE (fig. 5). There are apparently no hiatuses in this record, although, as in any sedimentary section, we cannot rule out the possibility of small stratigraphic gaps ( $\ll 100$  k.y.). The Bass River section is therefore one of the most stratigraphically complete records of the LPTM. Good to excellent preservation of foraminifera throughout the section has allowed the generation of a carbon and oxygen isotopic record through the LPTM providing a record of climate changes in a neritic setting. The location of Bass River in a neritic setting also allows examination of changes in the terrestrial climate through such proxies as clay mineralogy. In addition, identification of three planktonic foraminiferal taxa previously identified at ODP Site 865 [Kelly *et al.*, 1996] confirms the global distribution of these taxa which evidently evolved in response to environmental changes during the LPTM. Studies at Bass River have also allowed the documentation of a faunal turnover in calcareous nannoplankton at the LPTM (Aubry, unpublished data).

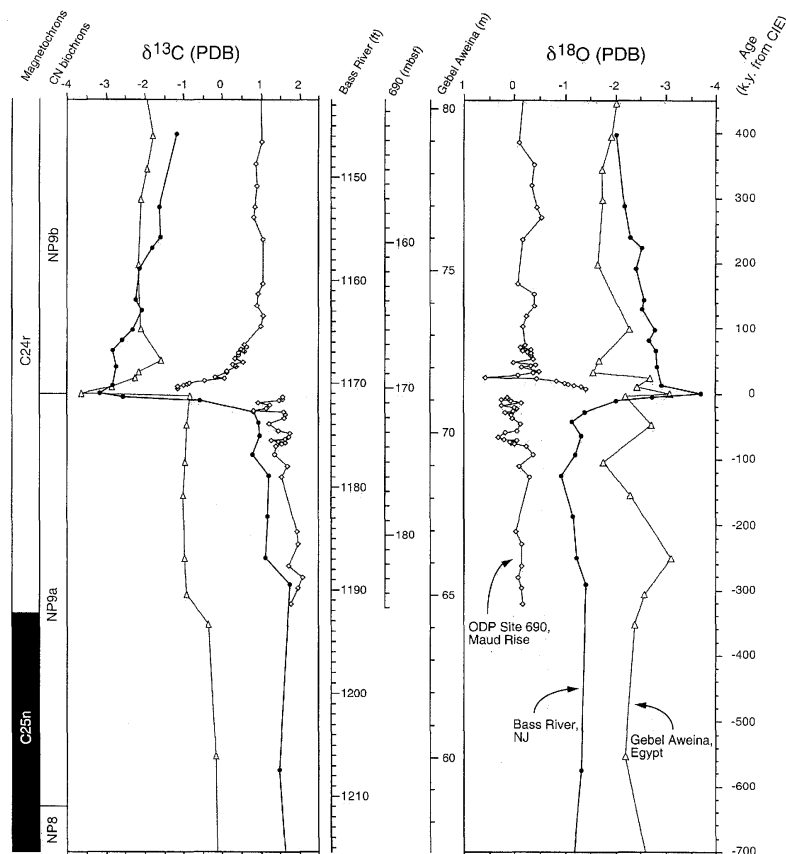


FIG. 8. – Comparison of isotope records from Bass River (this study), ODP Site 690 [Kennett and Stott, 1990, 1991], and Gebel Aweina [Schmitz *et al.*, 1996]. Bass River *Cibicidoides* spp. analyses are plotted as circles, ODP Site 690 *Nuttalides truempyi* analyses [Kennett and Stott, 1990, 1991] are plotted as diamonds, and Gebel Aweina *Lenticulina* spp. analyses [Schmitz *et al.*, 1996] are plotted as triangles. The records for ODP Site 690 and Gebel Aweina are plotted relative to Bass River using tie-points at the CIE and the base of Chron C24r for Site 690 [185.47 mbsf, Spiess, 1990] and the base of nanoplankton Zone NP9 for Gebel Aweina [58.6 m, Schmitz *et al.*, 1996]. Age model is based on a sedimentation rate of 19.3 m/m.y. at Bass River.

FIG. 8. – Comparaison des courbes isotopiques dans les coupes de Bass River (cette étude), du Site ODP 690 [Kennett and Stott, 1990, 1991], et du Gebel Aweina [Schmitz *et al.*, 1996]. Cercles : analyses sur *Cibicidoides* spp., Bass River; losanges : analyses sur *Nuttalides truempyi*, ODP Site 690 [Kennett and Stott, 1990, 1991]; triangles : analyses sur *Lenticulina* spp., Gebel Aweina [Schmitz *et al.*, 1996]. Les courbes du Site ODP 690 et du Gebel Aweina sont établies par rapport à la courbe de Bass River, en utilisant les points de repères que constituent le CIE et la base du chron C24r au Site 690 [185,47 mbsf, Spiess, 1990], et la base de la zone NP9 dans la coupe du Gebel Aweina [58,6 m, Schmitz *et al.*, 1996]. Ages établis sur la base d'un taux de sédimentation de 19,3 m/m.y.

### Implications for New Jersey lithostratigraphy

The Bass River section provides an opportunity to reevaluate previous records of the upper Paleocene-lower Eocene from Clayton and Island Beach, NJ. As in the Clayton section, the CIE at Bass River marks the transition between the slightly sandy, glauconitic lower and clay-rich upper portions of the uppermost Paleocene unnamed unit (compare figures 2, 5 in this study with figure 3 of Gibson *et al.* [1993]). Based on a lithologic and stratigraphic comparison between Island Beach, Bass River, and Clayton [Cramer, 1998], it is clear that the section at Island Beach does not contain the upper, sand-barren portion of this unit (compare figure 2 of Liu *et al.* [1997]). Pak *et al.* [1997] tentatively correlated an isotope excursion in the lowermost Manasquan sediment at Island Beach with the CIE. Aubry *et al.* (manuscript in prep.) have demonstrated that the unconformity at Island Beach at the level of the excursion noted by Pak *et al.* [1997] separates nanofossil Subzone NP9a from

Subzone NP10a with only ~1 ft (0.3 m) representing Subzone NP9b just below the unconformity. This implies that the isotopic record from Island Beach is a juxtaposition of the very initial portion of the CIE with lighter Lower Eocene isotopic values.

### Isotope shift

The shift in isotopic values at Bass River is large in comparison with other records. In general shape, the carbon isotope record is similar to records from Gebel Aweina and Gebel Duwi, Egypt [Schmitz *et al.*, 1996] (fig. 8), showing the persistence of low  $\delta^{13}\text{C}$  values for > 200-400 k.y. after the shift; in contrast, deep-sea records show a nearly full recovery to pre-excursion values within < 100 k.y. [e.g. Site 690 (fig. 8) : Kennett and Stott, 1990, 1991; Site 865 : Bralower *et al.*, 1995]. For this reason we characterize the isotopic record at Bass River as containing a "shift" in isotopic

values rather than an "excursion" which implies both a shift and a rapid recovery. This shift is correlative with the latest Paleocene CIE and occurred on similarly short timescales (< 10 k.y., fig. 6). It is likely that the persistence of low  $\delta^{13}\text{C}$  reflects differences in cycling of carbon between shelf environments and open-ocean environments.

The oxygen isotope record at Bass River also shows a persistence of low values above the excursion, similar to the pattern in carbon isotopes. Application of the paleotemperature equation assuming no ice volume [ $\delta_w = -1.2\text{‰}$ , Shackleton and Kennett, 1975] implies an increase in bottom water temperature on the shelf from  $\sim 14\text{ °C}$  pre-excursion to  $\sim 20\text{ °C}$  post-excursion with a peak temperature of  $25\text{ °C}$  at the CIE. This is most likely a maximum estimate, since increased precipitation could result in a significant decrease in  $\delta^{18}\text{O}$ . If the  $\sim -1.8\text{‰}$  shift in  $\delta^{18}\text{O}$  values between 1173.9 ft (357.81) and 1170.4 ft (356.74 m) was due entirely to freshwater input this would imply a decrease in salinity of  $\sim 8\text{‰}$  (assuming modern Delaware River, PA  $\delta^{18}\text{O}$  of  $\sim -8\text{‰}$ , Fairbanks [1982]) to  $\sim 12\text{‰}$  (assuming modern Roanoke River, NC  $\delta^{18}\text{O}$  of  $-5\text{‰}$ , Fairbanks [1982]). Realistically, the oxygen isotopic record from Bass River must reflect some combination of temperature and freshwater input. Warming (and increased precipitation) occurred in less than 20 k.y. with a slight rebound to cooler temperatures within  $\sim 10$  k.y. immediately following the CIE and gradual cooling continuing for at least the subsequent 300-500 k.y. (figs. 5, 6).

In contrast, neritic isotope records from Egypt show no significant excursion in oxygen (fig. 8) [Schmitz *et al.*, 1996]. Although the oxygen isotopic record from Egypt may reflect diagenetic alteration, one other published dataset supports the idea that tropical sea-surface temperatures may, in fact, have been stable through the LPTM [ODP Site 865, equatorial Pacific, Bralower *et al.*, 1995]. The record at Bass River demonstrates that warming did occur in shallow waters at relatively low latitudes. We speculate that equatorial records do not show warming during the LPTM due to enhanced poleward transport of heat through the production of warm intermediate/deep waters in equatorial regions.

This speculation is not consistent with Schmitz *et al.* [1996], who rejected the hypothesis of production of warm deep water in the Egypt-Israel portion of the Tethys during the LPTM. Their conclusion was based on the observation that  $\delta^{13}\text{C}$  at Gebel Aweina and Gebel Duwi, Egypt, was significantly more negative during the LPTM than  $\delta^{13}\text{C}$  at ODP Site 690 (fig. 8). In the deep sea, bottom water  $\delta^{13}\text{C}$  becomes more negative as it travels away from the source due to the breakdown of organic matter as it settles from the sea-surface and in the surface sediment. However, in neritic sections carbon isotopic values are potentially strongly influenced by terrigenous input having extremely negative  $\delta^{13}\text{C}$  relative to the marine environment. This is a likely explanation for the relatively low carbon isotopic values in the records from Egypt and Bass River.

### Clay mineralogy

The clay mineralogy of marine sediment has frequently been used as an indicator of terrestrial climate [*e.g.*, Biscaye, 1965; Griffin *et al.*, 1968; Robert and Chamley, 1987; Chamley, 1989; Robert and Chamley, 1991, 1992]. The three components typically emphasized are kaolinite, illite and smectite. Kaolinite and smectite are both products of intense chemical weathering on land. Smectite forms in poorly drained soils in humid regions while kaolinite develops in well-drained soils receiving high precipitation. Illite is the major constituent of immature soils which have undergone little chemical weathering and is characteristic

of cold or arid climates or areas of steep relief which interfere with soil formation and prevent extensive chemical weathering. In marine sediments the relative proportion of these minerals generally reflects their abundance in nearby terrestrial environments, although the absolute abundance of each can be affected by the processes which transport them to the marine environment [see Chamley, 1989, for full discussion].

It is possible that the observed increase in kaolinite content across the LPTM in the Bass River section is due to dilution by another clay component below  $\sim 1175$  ft (358.1 m) which is greatly reduced above  $\sim 1170$  ft (356.6 m). It is unlikely that this component is illite, since illite is ubiquitous in the marine environment and its abundance generally responds to dilution by other clays such as kaolinite or smectite [Biscaye, 1965]. There could be a glauconite component contributing to the observed intensity of illite-smectite peaks. The disappearance of glauconite sand at the level of the CIE is consistent with a reduction in clay-size glauconite. However, as discussed above, we find no evidence in the XRD traces to support significant presence of clay-sized glauconite in the section. We therefore conclude that the observed increase in kaolinite content reflects an actual increase in kaolinite sedimentation at the Bass River site during the latest Paleocene.

In the Bass River section there is a very abrupt increase in kaolinite deposition relative to illite/smectite occurring in < 20 k.y. coincident with the CIE (fig. 6). Thiry [1998] has noted that it may take as long as one to several million years for the clay minerals in soil profiles to equilibrate with climatic conditions. Therefore, the increase in kaolinite in the Bass River section is much too rapid to reflect production of kaolinite in the terrestrial environment and must reflect erosion of previously deposited clays. The likely source of kaolinite in the Bass River section is erosion of exposed Upper Cretaceous deposits in the New Jersey Coastal Plain. In the Cenomanian Raritan Formation, deposited subaerially in a flooding plain and mangrove swamp environment, kaolinite content can reach > 90% and is generally  $\sim 60\%$  of the clay assemblage [Owens and Sohl, 1969]. Kaolinite remains a significant component of subaerial and marine deposits throughout the remainder of the Cretaceous. It is likely that erosion of these Upper Cretaceous deposits was ongoing in the late Paleocene and that variations in kaolinite content in the Bass River section must be due either to changes in the proximity of the Bass River site to the kaolinite source and/or to sorting mechanisms operating during transport and deposition of clay minerals.

Changes in proximity to source can be effected either by movement of a point source (river mouth) or by water depth changes affecting the distance from shore. Variation due to movement of a point source would cause very localized patterns of clay sedimentation. The similarity in shape and timing of the increase in kaolinite at Bass River (this study) and Clayton [Gibson *et al.*, 1993; Thomas *et al.*, 1997] implies that the variations in clay sedimentation were not localized. Further investigation is needed to confirm that variations in clay sedimentation were ubiquitous throughout the New Jersey coastal plain, but the similarity of records at Bass River and Clayton is inconsistent with a point source for kaolinite.

It is also unlikely that the variation in clay minerals is due to water depth variations. In general, kaolinite content of marine sediments decreases with increasing distance from shore while smectite and mixed-layer clays increase [*e.g.*, Biscaye, 1965; Chamley, 1989]. The abrupt increase in kaolinite is therefore most consistent with a decrease in water depth and distance from shore. The elimination of

the terrigenous sand component at the level of the CIE (1171.1 ft, 356.95 m) together with benthic foraminiferal faunal evidence [Cramer, 1998] indicates an increase in water depth. There is a hint that kaolinite sedimentation covaries with sand in the interval below ~ 1180 ft (359.7 m), suggesting that kaolinite content may in fact be responding to minor changes in water depth in this interval, but the high kaolinite content above the CIE is clearly unrelated to water depth (fig. 5).

The rapid increase in kaolinite relative to illite/smectite coincident with the CIE is most likely due to environmental changes differentially affecting transportation of clay minerals. The  $\delta^{18}\text{O}$  record indicates an increase in temperature and a decrease in salinity due to increased precipitation at the CIE, persisting for the remainder of the section (1138.6-1171.8 ft, 347.05-357.17 m). An increase in precipitation is consistent with (1) an increase in terrigenous components such as kaolinite and/or (2) differential transport of kaolinite relative to illite/smectite into the neritic environment<sup>1</sup>. Whatever the sorting mechanism, we believe that the precise coincidence of the kaolinite peak with the carbon and oxygen isotopic shifts (both in New Jersey and high latitude ODP Site 690) [Robert and Kennett, 1992, 1994] and the lack of consistency between variations in kaolinite content and other sedimentologic parameters indicates a climatic control on kaolinite deposition during and after the LPTM.

## SUMMARY

Isotopic records from the upper Paleocene section at Bass River section show major negative shifts in  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  correlative with the latest Paleocene CIE recognized in many other marine and terrestrial sections. Integrated magneto- and nannofossil bio-stratigraphy demonstrates that this record is complete from near the base of Chron C25n to mid-Chron C24r and mid-Zone NP8 to upper Zone NP9b. Therefore, the Bass River section preserves a record of the LPTM in a neritic setting which is apparently as complete

<sup>1</sup> It has been suggested that kaolinite flocculates readily in relatively low salinity water and is deposited in estuarine environments while smectite and mixed layer clays are preferentially transported into the marine environment [Chamley, 1989].

as the deep sea record from ODP Site 690. The magnitude of the negative shift in carbon (4.0 ‰) and oxygen (2.3 ‰) isotopes is comparable to the record from ODP Site 690. The negative shift in  $\delta^{18}\text{O}$  indicates that rapid warming (and wetting) at relatively low latitudes (Bass River) as well as at high latitudes (ODP Site 690) occurred during the LPTM.

The isotope excursion is accompanied by a sharp increase in kaolinite content. The onset of the isotope excursion and the spike in kaolinite content are synchronous and occur within < 20 k.y. The increase in kaolinite correlative with the LPTM observed at ODP Site 690 [Robert and Kennett, 1994], Clayton, NJ [Gibson *et al.*, 1993], and in this study at Bass River, NJ appears to be a climate signal related to increased temperature and/or precipitation. The mechanism for producing this climatic response has not yet been determined.

The climate system in New Jersey apparently did not return to pre-LPTM conditions during the remainder of the Paleocene Epoch. Carbon and oxygen isotopes do not return to pre-excursion values after the excursion in the Bass River section, a pattern which is similar to the neritic records from Egypt [Schmitz *et al.*, 1996]. Persistence of low  $\delta^{13}\text{C}$  after the CIE may be a shallow-water phenomenon; however, a full explanation awaits further study and records from other neritic sites. Continued high kaolinite content and low oxygen isotope values after the excursion suggest that the climate in the neritic environment, at least in New Jersey, remained warmer and wetter for at least 300-500 k.y. after the excursion.

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