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SEDIMENTARY PATTERNS ACROSS THE PALEOCENE-EOCENE BOUNDARY IN THE ATLANTIC AND GULF COASTAL PLAINS OF THE UNITED STATES

by

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ABSTRACT

Fossiliferous clay and sand belonging to calcareous nannofossil Zones NP 9 (latest Paleocene) and NP 10 (earliest Eocene) are widespread in the U.S. Gulf and Atlantic Coastal Plains. Although the thickness of Zone NP 9-NP 10 strata is several times greater in the eastern Gulf Coastal Plain than in the Atlantic Coastal Plain, the hiatus that usually is present between these two zones in inner neritic strata is of longer duration in the Gulf Coast. The eastern Gulf Coastal Plain contains 2 thin and 2 thick, uppermost Paleocene (Zone NP 9) depositional sequences and 6 thin, lower Eocene (Zones NP 10-NP 11) sequences. Individual sequences are more difficult to distinguish in Atlantic Coastal Plain deposits, presumably because of a lower sedimentation rate, but 2 thin, uppermost Paleocene (Zone NP 9) and 6 thin, lower Eocene (Zones NP 10-NP 11) sequences are recognized. Although 6 sequences are found in Zones NP 10-NP 11 in both the Atlantic and Gulf Coastal Plains, the number of sequences in each NP zone differs in the 2 areas, which suggests that erosion has removed entire sequences in both areas.

There is continuous deposition across the Zone NP 9-NP 10 boundary only in the northern Atlantic Coastal Plain, where middle to outer neritic deposits record the presence of several latest Paleocene (upper Zone NP 9) events that occurred at approximately 55.2 Ma. These events include 1) a change from glauconitic sand to non-glauconitic clay, 2) a change in dominant clay mineralogy from smectite and illite to kaolinite, 3) a sharp decrease in the benthic foraminiferal species diversity along with a change from more normal marine assemblages to lower oxygen assemblages, 4) a large increase in the abundance of planktonic foraminifers, and 5) a significant increase in the number of first and last appearances of both calcareous nannofossil and sporomorph assemblages. These foraminiferal and sediment changes persist into the earliest Eocene (lowermost Zone NP 10) and there are no significant changes precisely at the Paleocene-Eocene (Zone NP 9-NP 10 boundary).

KEY WORDS

Stratigraphy, sequence analysis, calcareous nannofossils, foraminifera, U.S. Gulf, U.S. Atlantic, Paleocene, Eocene.

1. INTRODUCTION

Uppermost Paleocene and lowermost Eocene deposits are widespread in the eastern Gulf of Mexico and Atlantic Coastal Plains. The diverse and well-preserved calcareous microfossil assemblages that occur in these U.S. deposits can be correlated with global planktonic zonations. The microfossil assemblages also can be used for detailed paleoenvironmental interpretation of sediment deposited during the Paleocene-Eocene boundary interval, which is a time of global atmos-pheric, biologic, climatic, and oceanographic changes (Tjalsma & Lohmann, 1983 ; Miller *et al.*, 1987 ; Rea *et al.*, 1990 ; Thomas, 1989, 1990 ; Robert & Chamley, 1991 ; Kennett & Stott, 1991). The calcareous faunas and floras that are present in these strata are important to our understanding of North Atlantic shallow-water depositional environments and biostratigraphy near the Paleocene Eocene boundary ; they are particularly important for interpreting lowermost Eocene deposits because coeval sediments

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in northwestern Europe do not contain calcareous planktonic microfossils.

We will describe lithologic and biologic characteristics of uppermost Paleocene and lowermost Eocene deposits in the eastern Gulf of Mexico and Atlantic Coastal Plains, and then discuss the similarity both in nature and timing of the occurrence of several environmental changes found in uppermost Paleocene neritic deposits in the northern Atlantic Coastal Plain to those observed at or near this time in deep-sea deposits. These events, which previously have been recognized most commonly in high-latitude deep-oceanic environments, can now be extended into mid-latitude neritic environments in the western North Atlantic.

2. METHODS AND MATERIALS

Alabama contains the most extensive outcrops of upper Paleocene and lower Eocene fossiliferous marine strata in the eastern Gulf of Mexico Coastal Plain (Fig. 1), and we examined hundreds of samples from the eastern, central, and western parts of this state (Fig. 2). Exposures of marine strata of this age are much more limited in western Georgia and eastern Mississippi, and fewer samples were collected there. In the Atlantic Coastal Plain, the most extensive outcrops are in Virginia and Maryland (Fig. 3). Few calcareous, fossiliferous, upper Paleocene and lower Eocene outcrops occur in other Atlantic margin states.

Because of the limited number of coastal plain outcrops, we drilled continuously cored holes in selected areas of the Atlantic and eastern Gulf Coastal Plains. Some holes were sited in areas only slightly downbasin from the outcrop belt, where relatively shallow drilling of 100 m or less could penetrate entire formations. In addition, a few moderately deep holes to 200 m or slightly greater depths were drilled a considerable distance downbasin from the outcrop belt in the Atlantic Coastal Plain.

Most outcrop and corehole sections were extensively sampled for calcareous nannofossils at intervals of 0.3-1.0 m or less. For example, 168 samples were examined from the 33-m-thick composite section of the Bashi and Hatchetigbee Formations (calcareous nannofossil Zones NP 10-NP 11 of Martini, 1971) in Coffee County, Alabama. In the Solomons Island, Maryland, corehole, 56 samples were examined from the 96 m of strata of Zones NP 5 to NP 13. Biostratigraphic and paleoenvironmental analyses of foraminifers also were conducted on numerous samples. For example, 93 samples were examined from the Coffee County composite section, and 45 samples were examined from the Solomons Island corehole.

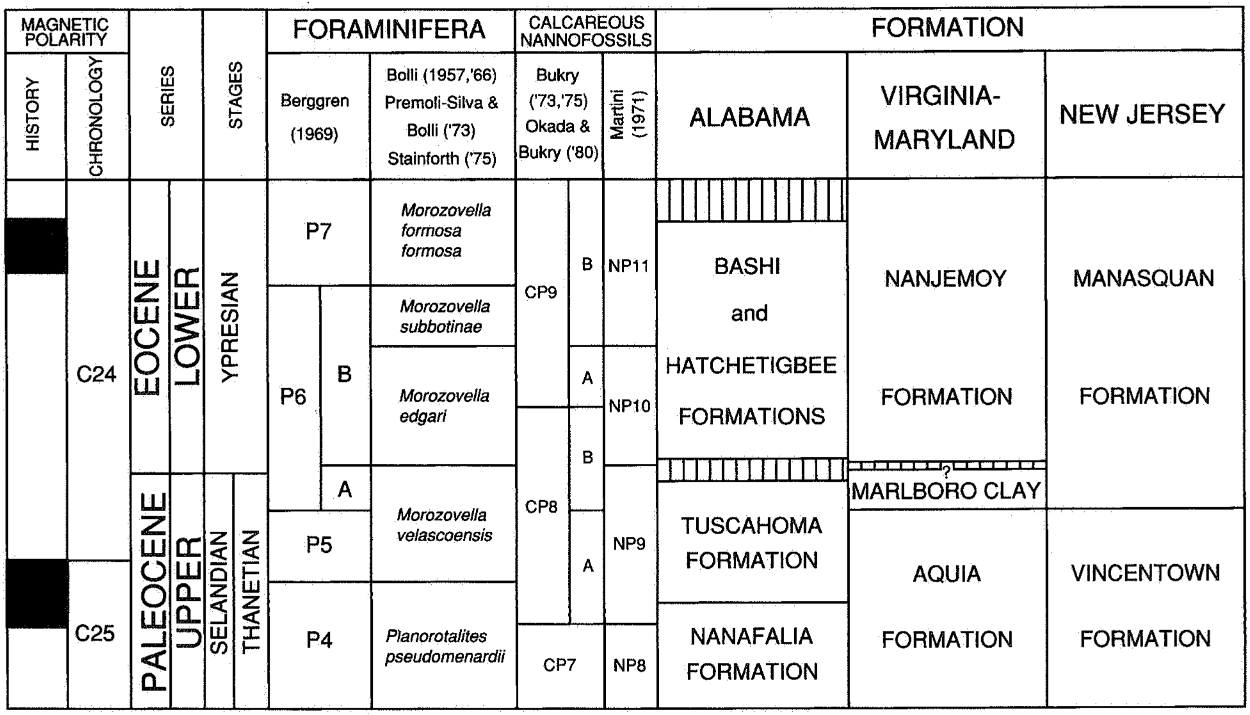

Figure 1. Correlation chart of upper Paleocene and lower Eocene formations in Alabama and the central and northern U.S. Atlantic Coastal Plain (time scale and zonation data from Berggren et al., 1985).

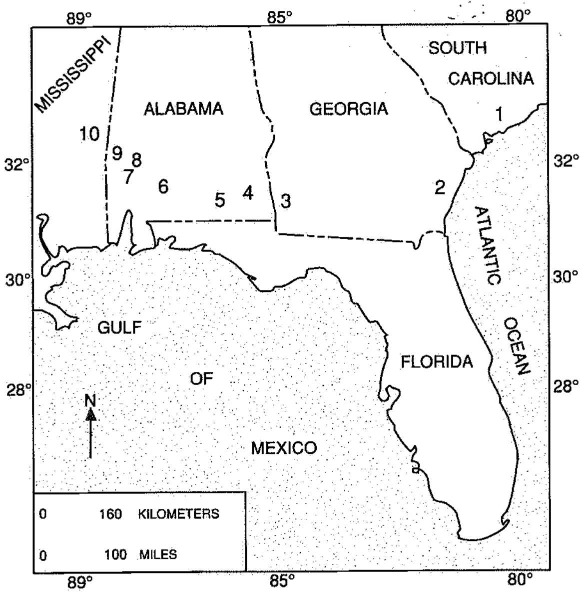

Figure 2. Map of localities in the SE U.S.: 1. Clubhouse Crossroads # 1 corehole ; 2. Davis-Hopkins corehole ; 3. Blakely ; 4. Ozark ; 5. Coffee County ; 6. Tunnel Springs ; 7. Hatchetigbee Bluff ; 8. Bashi Creek and Highway 69 ; 9. OSM No. 2 Wahalak corehole ; 10. Meridian.

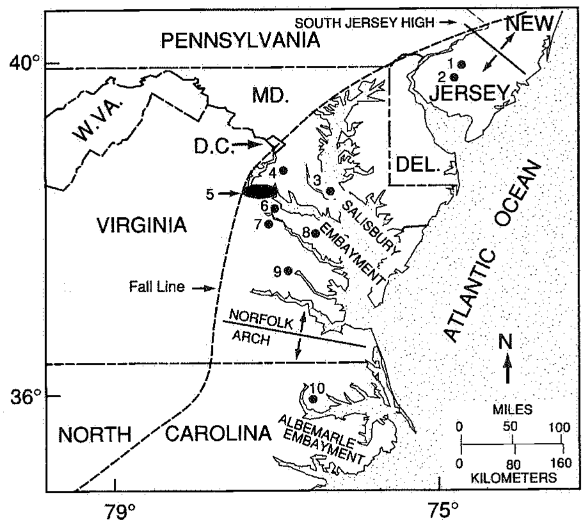

Figure 3. Map of localities in the central and northern U.S. Atlantic Coastal Plain: 1. GL 913, GL 915, and GL 917 drillholes ; 2. Clayton corehole ; 3. Solomons Island corehole ; 4. Waldorf corehole ; 5. Potomac River outcrop area ; 6. Oak Grove corehole ; 7. Loretto corehole ; 8. Haynesville corehole ; 9. Putney Mill corehole ; 10. Valhalla corehole.

To prevent contamination of corehole samples, calcareous nannofossil samples were taken from the center portion of freshly broken core surfaces. The outer surface of the core segments that were used for foraminiferal analysis was removed in order to eliminate drilling mud contamination. Foraminiferal data were derived from a plus-63 micron aliquot that contained 300-600 benthonic specimens.

3. PALEOCENE-EOCENE BOUNDARY

Although various placements have been proposed for the Paleocene-Eocene boundary (Berggren *et al.*, 1985), at the present time the most commonly used placement is at the calcareous nannofossil Zone NP 9-NP 10 boundary (Fig. 1). Aubry *et al.* (1988), however, based upon a re-examination of northwestern European sections, posi-tioned the Paleocene-Eocene boundary in the approximate middle of Zone NP 10 rather than at the base of this zone. The eastern U.S. sections contain a good calcareous fossil record for much of Zone NP 10, and they provide invaluable biostratigraphic and paleoenvironmental information on the Paleocene-Eocene boundary, whether it is placed at the base of Zone NP 10 or higher within this zone.

The Zone NP 9-NP 10 boundary was assigned an age of 57.8 Ma by Berggren *et al.* (1985) and 57.6 Ma by Aubry *et al.* (1988). Analysis of sanidine crystals from a North Sea basin ash layer within lower Zone NP 10 gave ages of 54.5 Ma (Swisher & Knox, 1991) or 55.1 Ma (Obradovich, 1988). Swisher and Knox extrapolated, on the basis of rates of sedimentation, to give an estimated age of approximately 55 Ma for the Zone NP 9-NP 10 boundary ; the use of the Obradovich age for this horizon within lower Zone NP 10 would place the zonal boundary at approximately 55.6 Ma. We used these younger ages for the Zone NP 9-NP 10 boundary to calculate the age of the events found in the Atlantic Coastal Plain.

Planktonic foraminiferal species that are used to delineate the Paleocene-Eocene boundary, such as *Morozovella velascoensis*, often have sporadic occurrences in shallow-water or non-tropical sections (THOMAS, 1989). Key planktonic foraminiferal species are rare in the shallow-water environments that are typical of the majority of eastern U.S. boundary interval sediments. Diagnostic calcareous nannofossil assemblages, however, usually are present in these deposits in the coastal plains, and they provide the best biostratigraphic control for eastern U.S. strata in the late Paleocene and early Eocene.

4. EASTERN GULF OF MEXICO COASTAL PLAIN

4.1. Regional patterns

Relatively high sedimentation rates were present in the eastern Gulf Coastal Plain during the late Paleocene (Zone NP 9) and early Eocene (Zones NP 10-NP 11). Rainwater (1968) considered that upper Paleocene and lower Eocene strata represented the time of greatest input of Tertiary deltaic sediments into the Gulf Coast.

Lithologic units that bracket the Paleocene-Eocene boundary in eastern Mississippi, Alabama, and western Georgia (Fig. 2) include the Tuscahoma Formation (Zone NP 9) and the Bashi and Hatchetigbee Formations (Zones NP 10 and NP 11) (Gibson *et al.*, 1982) (Fig. 1). These formations are considerably thicker in western Alabama, which is close to large Paleogene deltaic discharges in Mississippi (Galloway, 1968), than they are in eastern Alabama and western Georgia where sediment input was considerably lower bec-ause of the greater distance from the deltas.

The Tuscahoma Formation is 154 m thick in the OSM NO. 2 Wahalak corehole in westernmost Alabama (Fig. 2) (Mancini, 1981), but in easternmost Alabama it is commonly only 40-85 m thick in downbasin areas and only 18 m thick in upbasin areas (Gibson *et al.*, 1982). The Bashi and Hatchetigbee Formations are 76-84 m thick in western Alabama (Scott, 1972 ; Mancini, 1981). In easternmost Alabama, these formations reach a total thickness of 21 m in upbasin areas and are as little as 11 m thick in downbasin areas (Gibson *et al.*, 1982). In addition to thicker sections and presumably greater subsidence in western Alabama during this time interval, marine deposition within individual transgressive units, as determined from outcrop samples, apparently started earlier in the west. For example, the Tuscahoma Formation in western Alabama contains 4 marl units (Mancini & Oliver, 1981), while there are only 2 marl beds in eastern Alabama (Gibson *et al.*, 1982). The additional 2 marl units in western Alabama are considered to be older than either of the 2 eastern Alabama marl units that occur there at the base of the Tuscahoma section, and they probably represent earlier Zone NP 9 marine deposition in western Alabama. Based upon the calcareous nannofossil assemblages (see 4.5), slightly older strata may occur in the basal part of the Bashi Formation in western Alabama than in central and eastern Alabama.

Because eastern Alabama was a greater distance from the major deltaic influx that was occurring in Mississippi than was western Alabama, the formations in eastern Alabama are dominated by marine deposits and contain fewer deltaic deposits than those in western Alabama.The presence in eastern Alabama, both in outcrop and in the subsurface, of thinner, marine-dominated sections facilitates observation of the superpositional relationship of the several transgressive and regressive intervals within the Bashi and Hatchetigbee Formations. In western Alabama, the thin transgressive marine units are separated by thick deltaic intervals, and it is more difficult to observe the multiple transgressive-regressive cycles in the relatively thin outcrop sections found there. It also is necessary to drill significantly deeper coreholes in western Alabama than in eastern Alabama to penetrate the entire succession of cycles. In addition, the more highly dissected topography in easternmost Alabama typically exposes a considerably greater thickness of strata within this area's outcrops that allows the tracing of individual facies from upbasin to down-basin areas (Gibson, 1980a).

4.2. Uppermost Paleocene lithostratigraphy

The Tuscahoma Formation in western Alabama is composed of 4 relatively thin beds (up to 3 m thick) of shelly, glauconitic sand or marl, which are separated by much thicker (up to 30 m thick) intervening units composed of interlaminated fine sand, silt, and clay beds that commonly contain carbonaceous debris, cross-bedded sand, and local lignite beds as much as 2 m thick. Few, if any, shells occur in these thick intervening units. The lowest 2 shell beds, informally termed the Bear Creek marl and the Fatama marl (Mancini & Oliver, 1981), are of limited extent even in western Alabama. The shelly glauconitic sands of the 2 highest shell beds, the Greggs Landing Marl Member and the overlying Bells Landing Marl Member (Fig. 4), are widely distributed in western Alabama. Ineastern Alabama, only 2 marl beds occur in the Tuscahoma (Gibson, 1980a) ; they are considered to represent the upper 2 marl beds, the Greggs Landing Marl and Bells Landing Marl Members.

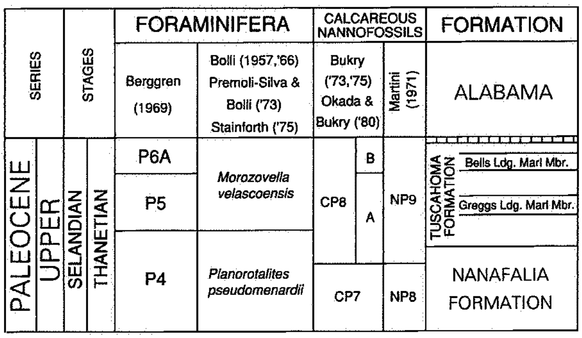

Figure 4. Correlation chart of upper Paleocene formations in Alabama (time scale and zonation data from Berggren et al., 1985).

Baum & Vail (1988) and Mancini & Tew (1991) applied sequence stratigraphic concepts to the Tuscahoma Formation, and the reader is referred to these articles for a more comprehensive discussion. In their papers, each of the 2 upper shelly, glauconitic sand units, the Greggs Landing Marl and Bells Landing Marl Members, was postulated to be a transgressive marine portion of a depositional sequence (the transgressive systems tract of Posamentier *et al.*, 1988), and they proposed that the top of each shell bed represented a condensed section that contains the surface of maximum sediment starvation. They considered the commonly carbonaceous interlaminated beds of sand, silt, and clay, the cross-bedded sand, and the lignitic strata that overlie each of the 2 marl beds to be progradational, highstand regressive deposits of each sequence (highstand systems tract of Posamentier *et al.*, 1988). The largest portion of each sequence consists of these progradational deposits. Each of the upper 2 sequences, those containing the Greggs Landing Marl and Bells Landing Marl Members, was considered to be a third-order cycle (Baum & Vail, 1988 ; Mancini & Tew, 1991).

Mancini & Tew (1991) considered the 2 lower Tuscahoma marl units and intervening non-calcareous beds to be the upper part of a third-order sequence that began earlier in the late Paleocene at the base of the Nanafalia Formation. Each of the 2 lower Tuscahoma marl units probably represents a significant depositional event within this third-order sequence, and each marl may represent a transgressive unit of a fourth-order sequence or parasequence.

One to 3 lignite seams, which may be up to 2 m thick (Scott, 1972), commonly occur at or near the top of the youngest Tuscahoma sequence. These lignite beds, located in the upper part of Zone NP 9 and in Zone P6a, probably are similar in age to those found in the Sparnacian beds in France.

Calcareous microfossils have not been recovered to date from the progradational, highstand regressive deposits of the Tuscahoma sequences. It is unknown at this time whether a large portion of these regressive deposits was deposited in marine environments and lacks calcareous fossils because of some component of the primary depositional environment or subsequent diagenetic removal, or whether these sediments were deposited in brackish-water environments in which few calcareous organisms lived. The upward change in paleoenvironments of the Tuscahoma from open marine to tidal flat and lagoon (Gibson, Edwards & Frederiksen, 1980) suggests that during the later part of Zone NP 9 the thick prograding wedge of sediments of the upper Tuscahoma caused the basinward migration of marine deposition to locations well south of the present-day eastern Gulf Coast outcrop belt of the Tuscahoma.

4.3. Uppermost Paleocene biostratigraphy

The Tuscahoma Formation can be dated by the use of several microfossil groups. Gibson *et al.* (1982) and Siesser (1983) examined calcareous nannofossils from the Tuscahoma. Gibson *et al.*, 1982 found *Discoaster multiradiatus* without *Tribrachiatus* spp. in all 4 marl beds, thus placing all 4 in Zone NP 9 (Fig. 1). Siesser (1983) also found *D. multiradiatus* in a sample from the Bells Landing Marl Member, the only marl bed he examined. Mancini & Oliver (1981) examined planktonic foraminifers from all 4 marl beds. The occurrence of *Morozovella subbotinae* in the Bells Landing Marl Member placed this unit in Zone P6a (Fig. 1) ; the presence of *M. velascoensis*, but the absence of *M. subbotinae*, in the Greggs Landing Marl Member placed this unit in Zone P5. The lower 2 Tuscahoma marls, the informally named Bear Creek and Fatama marls of Mancini & Oliver (1981), contain *Planorotaloides pseudomenardii*, and they were placed in Zone P4 by Mancini & Oliver (1981). Overlap of the calcareous nannofossil and planktonic foraminiferal ranges places these lower 2 marl beds into uppermost Zone P4 and lowermost Zone NP 9 (Fig. 1). These ages place the entire marine calcareous part of the Tuscahoma Formation in Zone NP 9. The uppermost part of the Tuscahoma Formation, the highstand regressive deposits above the Bells Landing Marl Member, has not yielded calcareous microfossils. However, there are brackish to fresh water palynomorphs of late Paleocene affinities (Frederiksen *et al.*, 1982) in the uppermost beds of the Tuscahoma ; this suggests that the entire Tuscahoma is of late Paleocene age. The sporomorph assemblage from the uppermost beds of the Tuscahoma does not contain some species that are present in the uppermost part of Zone NP 9 strata in Virginia ; this absence suggests that strata equivalent to this youngest part of Zone NP 9 in Virginia are absent in Alabama (Fig. 1) (Frederiksen *et al.*, 1982).

4.4. Lowermost Eocene lithostratigraphy

A fossiliferous, glauconitic, silty, fine-grained sand unit, the Bashi Marl Member of the Hatchetigbee Formation of Toulmin (1977), commonly overlies the Tuscahoma Formation and may reach a thickness of 6-8 m, but commonly it is only 1-3 m thick. The Bashi Marl is similar lithologically to the underlying Greggs Landing and Bells Landing Marl Members of the Tuscahoma Formation. Toulmin placed the Bashi Marl at the base of the Hatchetigbee Formation, and he considered it to be overlain by interlaminated sand, silt, and clay beds and cross-bedded sand of the unnamed upper member of the Hatchetigbee Formation ; the latter deposits are similar lithologically to those found in the progradational, highstand regressive deposits of the Tuscahoma sequences (Toulmin, 1977). Thus, we find a continuation into the early Eocene of the latest Paleocene sedimentary patterns, and it is not surprising that the Bashi Marl has been 1) grouped with beds that now compose the upper part of the underlying Tuscahoma Formation (Smith *et al.*, 1894 ; Adams *et al.*, 1926), 2) considered to be a basal member of the overlying Hatchetigbee Formation (Toulmin, 1977), and 3) considered to be a separate formation that includes some beds previously placed in the unnamed upper member of the Hatchetigbee Formation (Gibson, 1983). Toulmin (1967, Fig. 6), illustrated the unnamed upper member of the Hatchetigbee as a thick detrital unit that overlies the Bashi in upbasin areas. He considered that in some areas, such as in eastern Alabama, strata equivalent to the unnamed upper member did not exist and that only the thin lower member, the marine Bashi Marl, was present. Gibson (1983) considered that the upper marine strata of the Bashi in down-basin areas in eastern Alabama are equivalent in age to the Hatchetigbee in upbasin areas as discussed below.

The Hatchetigbee Formation, as redefined by Gibson (1983), is composed dominantly of massive or cross-bedded, non-calcareous, non-glauconitic, very fine to fine-grained sand, interlaminated very fine sand, silt, and clay beds, and thickly bedded or massive clay. Carbonaceous debris is abundant in the interlaminated sequences in western Alabama (Scott, 1972), but it is less abundant in eastern Alabama and western Georgia. At a few localities, the beds have either thin shell lenses with a low-diversity molluscan assemblage, or scattered shells. Small amounts of glauconite are present in some of the sand beds.

Toulmin (1977) and Mancini & Tew (1991) considered their Bashi Marl Member of the Hatchetigbee Formation to be a single marine transgressive unit that is overlain by a single regressive complex of shallow-marine to marginal-marine beds (termed the unnamed upper member) of the Hatchetigbee Formation. This concept places the entire Bashi and Hatchetigbee units into 1 transgressive-regressive cycle, which Mancini & Tew (1991) considered to be a third-order sequence of Baum & Vail (1988).

In contrast, Baum & Vail (1988) placed the Bashi and Hatchetigbee strata into 2 third-order sequences. However, they provided no information concerning the strata or exposures upon which they based this subdivision. As discussed below (see 4.5), the proposed age for the younger of their 2 third-order sequences in the Hatchetigbee does not agree with published biostratigraphic data of Gibson *et al.* (1982) and Bybell & Gibson (1985).

During mapping of Paleocene and Eocene strata in eastern Alabama from 1977 to 1982, Gibson found, in addition to the basal shelly, glauconitic sand or marl bed of the Bashi Marl Member of Toulmin, several additional prominent shelly, glauconitic sand beds, similar or identical lithologically to the Bashi, scattered through the overlying unnamed member of the Hatchetigbee (Gibson *et al.,* 1982 ; Gibson, 1983). These multiple shelly, glauconitic sand or marl beds in the upper member of the Hatchetigbee were observed in exposures and coreholes across Alabama. Multiple marl beds in this interval are reported as far west as the vicinity of Meridian in easternmost Mississippi (Fig. 2) where Priddy (1961, pl. 1) noted that his Zone 4 of the Wilcox Group (Bashi marl) had 4 marls or marlstones interbedded with silts, clays, and lignites. Smith *et al.* (1894) is the earliest report of multiple marl beds in the Bashi and Hatchetigbee interval ; they reported 3-4 marl beds in both eastern and western Alabama.

In eastern Alabama, the marl beds are more closely spaced because the inter-vening less marine lithologies are thin ; this is a result of their greater distance from the major deltaic input into the basin. In southwestern Georgia, along the Chattahoochee River near Blakely (Fig. 2), the lowest 2 shelly, glauconitic sand beds coalesce to form one thick bed of Bashi lithology (Gibson & Bybell, 1981, section 99, Fig. 3). Upbasin (northward) from this area, the marl beds thin and separate, and there is a progressively greater thickness of Hatchetigbee lithology between the 2 marls. The shelly, glauconitic, very fine grained sand bed(s) reflect shallowing marine conditions to the north where they change to massive, fine-grained sand with thin lenses or pockets of shell and then to fine- to medium-grained, pebbly, clay clast-bearing cross-bedded sand in the farthest upbasin areas located at or near the Bashi shoreline (Gibson, 1980a).

Our studies reveal that in eastern Alabama there are 6 marl beds that commonly grade up into a carbonaceous clay, which is disconformably separated from the next overlying marl bed. We have designated each shelly, glauconitic sand bed and the commonly present overlying carbonaceous clay as one depositional cycle. All 6 cycles (Fig. 5) occur in superposition in Coffee County, Alabama (Fig. 2).

The upper and lower surfaces of each cycle are highly burrowed ; the burrows are filled with shelly glauconitic sand and extend as much as 1 meter into the underlying carbonaceous clay, and there may be as much as 1 meter of erosional relief to this disconformity. All or part of the upper carbonaceous clay bed in a cycle may be missing due to erosion before the deposition of the overlying cycle, and in some cases, even the lower, glauconitic sand bed of the cycle is partially eroded away. For example, the marine beds of the fifth cycle are truncated at a level where the beds still represent relatively deep-water, inner middle neritic environments of the transgression. Further upbasin, the sediments occur at higher elevations, and erosional truncation here may completely remove from a few to as many as 5 of the 6 cycles. Some of this erosion was post-Eocene in age, and it removed much or all of the lower Eocene deposits from large areas in a manner similar to that described for the Atlantic Coastal Plain (Gibson *et al.,* 1992). However, portions of 1 or more sequences also were eroded during the early Eocene. For example, only the lowest 3 early Eocene cycles are preserved at Ozark, Alabama

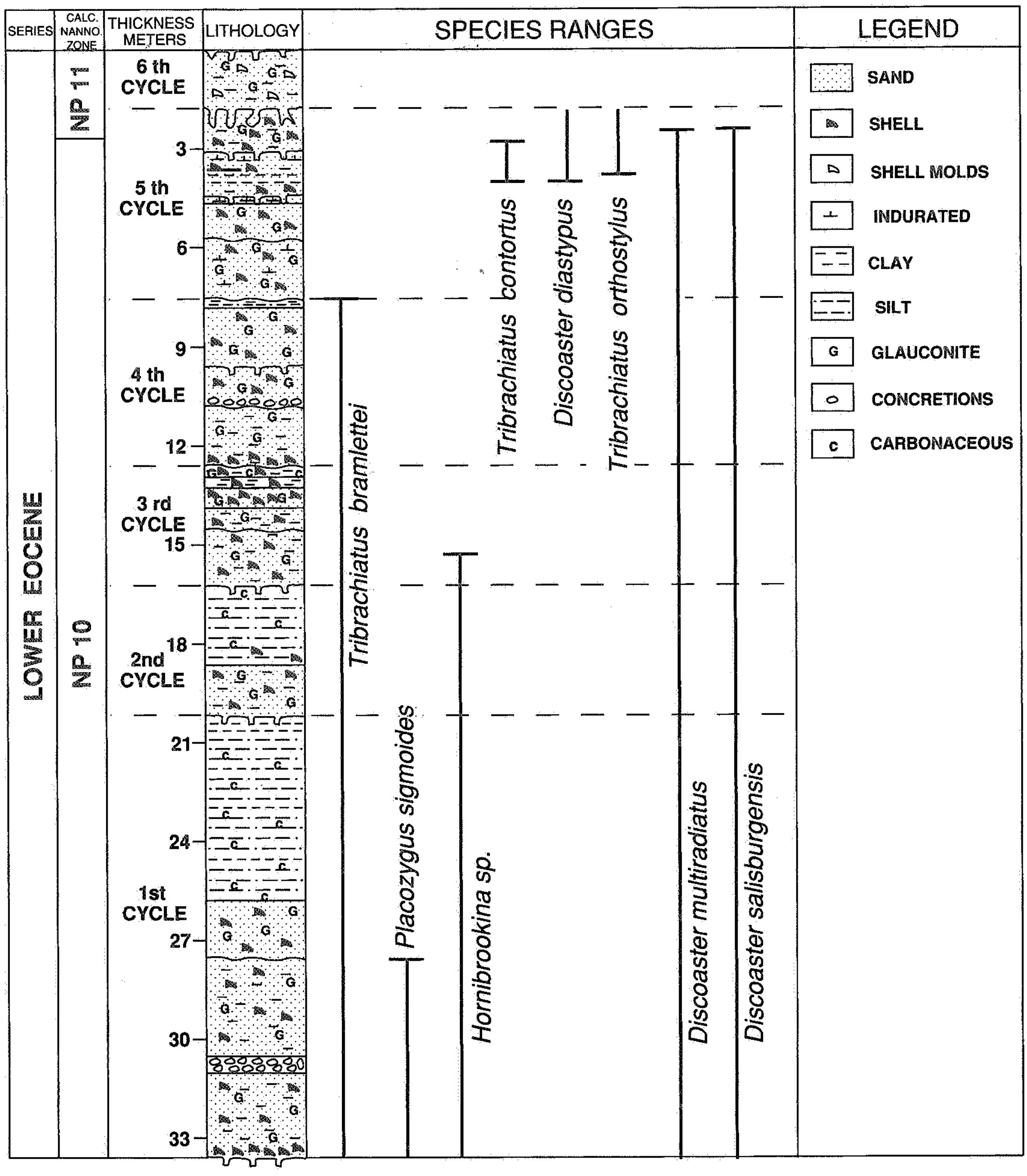

Figure 5. Composite section in Coffee County, Alabama, showing 6 cycles of the lower Eocene Bashi and Hatchetigbee Formations and the ranges of diagnostic calcareous nannofossil taxa.

(Fig. 2) (Gibson, 1988, Fig. 16) ; the upper 3 cycles were removed before deposition of the overlying Tallahatta Formation in the late early Eocene (Bybell & Gibson, 1985). Only the lowest 2 cycles are preserved in outcrop across much of easternmost Alabama. The significantly greater thickness of the lowest of the 6 cycles (by a 3-fold factor, Fig. 5), in combination with its relatively lower outcrop elevation and thus lower erosional gradient, accounts for this being the most commonly preserved cycle. The basal marl bed of this lowest cycle is the Bashi Marl Member of Toulmin.

Based upon the foraminiferal assemblages, the marl bed in the upper part of the fifth cycle appears to represent the deepest water deposition and highest sea level of the 6 cycles. This cycle was removed by erosion in almost all upbasin areas and is truncated even in downbasin outcrop areas in Alabama by erosion that occurred later in Zone NP 11 or early in Zone NP 12 before deposition of the overlying Tallahatta Formation. Thus, the original upbasin extent and maximum water depth represented by this cycle cannot be determined with certainty. Paleobathymetry of the fifth cycle, the upper part of which is placed in Zone NP 11, is of particular interest in the study of global sea le vels. As discussed below, Zone NP 11 deposits in the Atlantic Coastal Plain represent the deepest water early Eocene deposition in Virginia and Maryland. Zone NP 11 also is the time of the initial appearance of abundant planktonic foraminifers and calcareous nannofossils in lower Eocene strata in southern England and northern France (Wright, 1972 ; King, 1981). The occurrence of calcareous planktonic microfossils in these European strata suggests that higher sea levels, and thus more open circulation of Atlantic Ocean water into areas bordering the North Sea, were present for the first time in the early Eocene in Zone NP 11. Additional information on the time of deposition and the paleobathymetry of Zone NP 11 strata in the 2 a reas will determine if the similarity is due to local or global causes.

The 4 lower cycles of the Bashi-Hatchetigbee Formations are placed in the lower half of Zone NP 10 on the basis of the calcareous nannofossil assemblages (see 4.5). These 4 cycles represent a well-preserved calcareous record of the lower half of Zone NP 10. The upper half of Zone NP 10 and all of Zone NP 11 are less completely recorded here. The fifth cycle is placed in the upper half of Zone NP 10, and its uppermost 1 m belongs in the lower part of Zone NP 11. Because only the upper part of the fifth cycle plus the thin sixth cycle are placed in Zone NP 11, and because the lower part of the overlying Tallahatta Formation is placed in Zone NP 12 (Bybell & Gibson, 1985), it is likely that a considerable part of the total time of Zone NP 11 is not represented in Bashi and Hatchetigbee strata examined to date. All 6 cycles are not present within adjacent parts of Alabama, which indicates erosional removal of cycles as discussed earlier. The general paucity of cycles in the upper part of Zone NP 10 and Zone NP 11 in the eastern Gulf Coastal Plain is attributed primarily to erosional removal similar to that documented in the Atlantic Coastal Plain (Gibson *et al.,* 1992).

Zone NP 10 has a duration of about 1 my according to Berggren *et al.* (1985) and Haq *et al.* (1988) and 1.2 my according to Aubry *et al.* (1988). Because of evolutionary differences in the microfauna and microflora from cycle to cycle, at least some of the disconformities separating the cycles appear to represent a significant period of time within Zone NP 10. The sediments preserved in these 6 cycles are presumed to represent 100,000 year cycles.

The lowermost transgressive shelly, glauconitic sand of the Bashi Marl Member of Toulmin forms the initial deposit above the disconformity at the top of the Tuscahoma in all outcrop and subsurface sections in Alabama and western Georgia. The flat to very slightly undulating surface at the base of the Bashi appears to represent a marine planation surface. Recently, Ingram (1991) proposed that sand deposits, which occur in channels below the shelly glauconitic sand of the Bashi and above the Tuscahoma Formation at Meridian, Mississippi (Fig. 2), represent the oldest deposits of the Bashi Marl Member of Toulmin's sequence. Ingram considered that this lower, non-calcareous sand was deposited in a channeled surface (incised valley-fill, lowstand deposits of Baum & Vail, 1988) that developed during a sea-level lowstand ; the overlying shelly, glauconitic sand of the Bashi represents a subsequent marine transgression of the sequence that later reached this area and scoured and truncated the lowstand deposits.

Sedimentary characteristics and foraminiferal assemblages from the shelly, glauconitic sand that forms the lower part of each cycle suggest relatively low depositional rates in inner to middle neritic environments with water depths of 30-100 m (Gibson, 1988). Abundant planktonic foraminifers, diverse calcareous nanno-fossil assemblages, and diverse benthonic foraminiferal assemblages, which are dominated by *Cibicides, Cibicidoides, Hanzawaia*, and *Valvulinaria*, suggest oxygenated, normal salinity waters for these shelly glauconitic sand beds.

Carbonaceous, silty clay deposits, which form the upper part of most of the lower Eocene cycles, suggest that large volumes of clay and carbonaceous debris were introduced into the basin at this time. These sediments were carried by large volumes of river outflow from the ancestral Mississippi and other river systems that discharged into this area. Calcareous nannofossils generally are absent from these beds, presumably because the surface water wedge was highly turbid and had lowered salinity. Planktonic foraminifers commonly occur in the clay beds, although in somewhat reduced numbers. Benthonic foraminifers are abundant and diverse in most clay beds in downbasin localities. However, the assemblages differ from those found in the shelly sands ; assemblages in the clay are dominated by *Epistominella* spp., *Bulimina* spp., *Bolivina* spp., *Uvigerina* spp., *Pararotalia* spp., and *Pulsiphonina prima*. The benthonic assemblage and the presence of abundant carbonaceous debris in the sediments suggest that lowered oxygen and higher organic carbon conditions were present in the marine bottom waters.

The foraminiferal assemblages in some of the clay strata, particularly those in the upper part of the first cycle, suggest deposition in inner to inner middle neritic environments ; these water depths are similar to those suggested for the underlying shelly glauconitic sand beds in this cycle. The decrease in grain size from glauconitic fine sand below to carbonaceous clay higher in each cycle, while the paleobathymetry remained essentially unchanged, suggests that changes occurred in the transport system and/or in the type of weathering in the source area. These sedimentary changes and probable climatic changes occurred within the apparent 100,000 year periodicity of each of the cycles ; these changes may be the result of shifts in the global climate belts similar to those proposed by Perlmutter & Matthews (1989).

4.5. Lower Eocene biostratigraphy

Based on examination of the planktonic foraminiferal assemblages, Berggren (1965), Gibson (1980b), and Mancini (1981) placed the Bashi Marl Member of Toulmin in the earliest Eocene *Morozovella subbotinae* Interval Zone of Stainforth *et al.* (1975) or the *Morozovella edgari* Zone (lower Zone P6b) of Berggren *et al.* (1985) (Fig. 1). The presence of *M. acuta* in the planktonic assemblage, along with *M. subbotinae, M. wilcoxensis*, and *Pseudohastigerina wilcoxensis* (Gibson, 1980b ; Mancini, 1981), indicates that the Bashi Marl Member of Toulmin should be placed in the lower part of this zone.

Calcareous nannofossils provide a more detailed biostratigraphic subdivision of the 6 cycles. Figure 5 gives the ranges of some key calcareous nannofossil species from the most extensively sampled and most complete section yet known of the 6 Bashi cycles, in Coffee County, Alabama (Fig. 2). The lower 4 cycles contain *Tribrachiatus bramlettei*, a species restricted to the lower half of Zone NP 10 (Perch-Nielsen, 1985) (Fig. 6), which indicates that these 4 Bashi and Hatchetigbee cycles are in the lower half of Zone NP 10. A more complete depositional record is present for this time interval than for the upper half of Zone NP 10 and Zone NP 11 where only 2 relatively thin cycles are preserved. The lower part of the fifth cycle is a slightly to moderately indurated sand, and few calcareous microfossils were recovered from this part. The upper part of the fifth cycle contains in ascending order : 1) *Tribrachiatus contortus* only (upper half of Zone NP 10), 2) *T. contortus* and *Tribrachiatus orthostylus* (uppermost part of Zone NP 10), and 3) only *T. orthostylus* (Zone NP 11). Based upon the thin section in which this succession takes place, it is assumed that very low sedimentation rates were present in eastern Alabama at this time ; this part of the section probably represents a condensed interval deposited during a time of high sea level. The lower, barren part of the fifth cycle represents an unknown amount of time in the upper half of Zone NP 10. The upper part of the fifth cycle records part of upper Zone NP 10 and the transition from upper Zone NP 10 into lower NP 11. The sixth cycle is presently non-calcareous, but it does contain molluscan molds that signify a marine origin. Calcareous fossils presumably were dissolved from this bed during formation of the disconformity between it and the overlying beds of the Tallahatta Formation, which belong to Zone NP 12 (Bybell & Gibson, 1985).

Tribrachiatus bramlettei occurs in the basal 0.2 m of the Bashi Marl Member of Toulmin in numerous localities in eastern and central Alabama. This species also occurs in samples above the lower 0.3 m of the Bashi Marl Member of Toulmin from the Bashi type locality on Alabama Highway 69 just south of Bashi Creek (Fig. 2). A sample from the lowest 0.3 m of the Bashi there, however, does not contain *T. bramlettei.* The assemblage in this sample is poorly preserved and contains only 3 species. Well-preserved assemblages that contain 18-30 species occur in samples taken 1-3 m above the basal sample at this outcrop. A probable cause for the preservational and diversity difference between the lowest sample and the higher ones is that the more permeable sand bed of the Bashi rests upon an underlying clay bed of the Tuscahoma ; water percolating downward through the Bashi moved laterally at its base because of the confining nature of the underlying Tuscahoma strata and dissolved many of the calcareous microfossils from the lowermost part of the bed.

Although preservational conditions may have caused the absence of *T. bramlettei* from the lowest 0.3 m of the Bashi, an indeterminate species of *Fasciculithus* is present in the lowest sample. This genus is only present in the Bashi in Alabama at this one locality on Highway 69. Several species of *Fasciculithus* extend up into the lower part of Zone NP 10 (Perch-Nielsen, 1985). We also found that several species of *Fasciculithus* commonly occur with *Tribrachiatus bramlettei* in Zone NP 10 in the lower 3-5 m of the Nanjemoy and Manasquan Formations in the middle and northern Atlantic Coastal Plain (see 5.3.3). The presence of *Fasciculithus* in the lowest part of this exposure may indicate that older sediments occur here than in central and eastern Alabama where *Fasciculithus* is absent. This occurrence also suggests that early Eocene deposition began slightly later in Zone NP 10 in Alabama (above the LAD of *Fasciculithus*) than in the NE U.S. where *Fasciculithus* consistently is present.

Siesser (1983) placed the Bashi Marl Member of Toulmin in Zones NP 9 and NP 10. Although he considered all 4 of his samples to be in Zone NP 9, the criteria that he gave for this zonal placement (the co-occurrence of *Discoaster multiradiatus* and *D. mohleri*) are found in only 1 of his 4 samples as shown in his Table 5. Two of his samples contain a meager assemblage of 2 and 4 long-ranging species, and they do not contain either of his specified zonal species. The third sample, A21b, from Ozark, Alabama, (this sample is not from the Bashi Marl Member of Toulmin ; the shelly glauconitic sand bed that outcrops at this locality is the lower part of the second cycle of the Bashi, see Gibson, 1988, Fig. 16), contains only *D. multiradiatus.* Although this species' FAD is at the base of Zone NP 9, its range extends up into the lower part of Zone NP 11 (Perch-Nielsen, 1985) (Fig. 6). His fourth sample, A46a, from the Bashi type section on Highway 69 just south of Bashi Creek, contains both of his critical species as well as *Fasciculithus involutus.* However, as discus-sed above, *F. involutus* ranges up

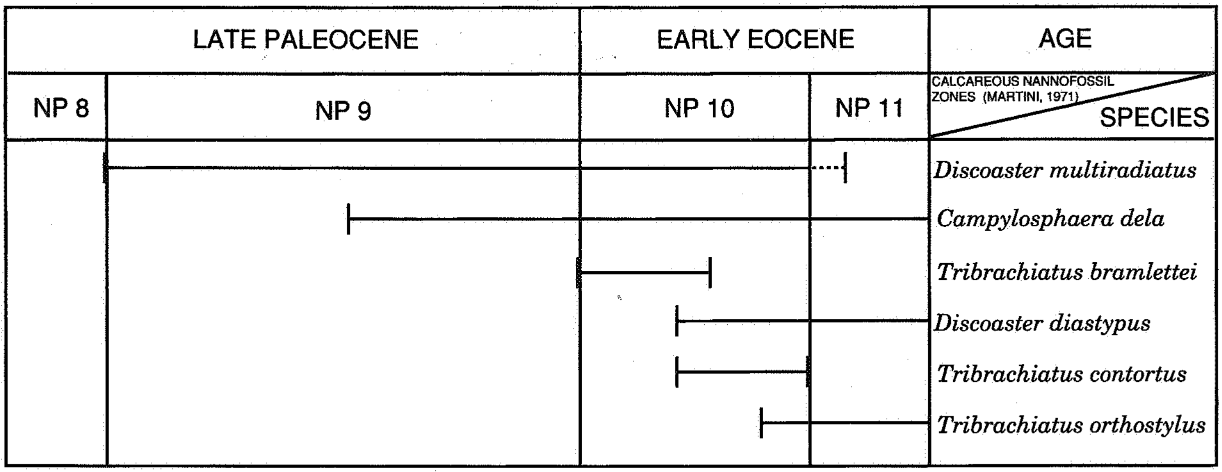

Figure 6. Ranges of diagnostic late Paleocene and early Eocene calcareous nannofossil taxa (from Perch-Nielsen, 1985).

into the lower part of Zone NP 10 (Perch-Nielsen, 1985), and *D. multiradiatus* ranges up into Zone NP 11. The illustrated specimen of his other diagnostic species, *D. mohleri* (Siesser, 1983, Fig. 15F), contains a stem, and appears to be similar to specimens of *Discoaster* aff. *D. mohleri* that we find together with *T. bramlettei* in lower Zone NP 10. Thus, we find no evidence that any of the Bashi can be placed in Zone NP 9.

As discussed above, all 6 cycles of the Bashi and Hatchetigbee Formations are placed in Zones NP 10 and NP 11. Bybell & Gibson (1985) demonstrated that the overlying Tallahatta Formation belongs in Zones NP 12 to NP 14. Baum & Vail (1988), however, placed the lower of their 2 Hatchetigbee sequences in Zones NP 10 and NP 11 and their upper sequence in Zones NP 12 and NP 13, which would overlap the Tallahatta. The location of their sections and the biostratigraphic data for these placements were not given. The lowest 2 cycles of the Bashi and Hatchetigbee Formations are the most widely exposed geographically. Accurate dating of the second cycle is difficult in many places as the calcareous microfossils commonly are removed by dissolution from the most accessible outcrops. We consider it more likely that the boundary between Baum & Vail's lower and upper Hatchetigbee sequence is located at the disconformity between the first and second cycle ; both of these cycles, however, belong in the lower part of Zone NP 10. The Tallahatta Formation contains the only strata in Alabama belonging to Zones NP 12 and NP 13.

Six depositional cycles occur in the Hampshire basin in southern England (Chris King, personal commun., 1991) in a time interval similar to that represented by the 6 Bashi and Hatchetigbee cycles in Alabama (from early Zone NP 10 to late Zone NP 11). It is unknown at the present time whether any of the 6 cycles coincide within the 2 different areas. Precise correlation between the 2 areas is difficult because Zone NP 10 strata in England contain neither calcareous nannofossils nor planktonic foraminifers and because dinoflagellate floras appear to have provincial differences between northwestern Europe and the southeastern U.S. during much of this time.

The Hampshire basin in England is located in a different global cyclostratigraphic belt than is Alabama, according to the climatic zonation of Matthews & Perlmutter (in press, Fig. 7). In their model, climatic patterns in England and Alabama, as expressed in the sedimentary record, should differ because they originated in different climatic zones. Thus, it is necessary to establish accurate age relationships among the 6 cycles in the 2 areas, as well as of the respective patterns of sedimentary change, in order to determine the relative importance of sea-level changes and climatic variations in the development of the Bashi cycles.

5. ATLANTIC COASTAL PLAIN

The thickest Cenozoic sediments that occur on land along the U.S. Atlantic margin were deposited in several embayments that resulted from downwarps in crystal-line basement rocks of Precambrian and early Paleozoic age (Fig. 7). The embayments are separated by arches or highs that are considered to be the result of flexures or faults in the underlying basement. The Cenozoic sediments preserved on the basement highs are much thinner and more discontinuous than those found in the adjacent embayments. In outcrop, the present day Fall Line (Fig. 7) marks the change from Precambrian and Paleozoic schist, gneiss, and granite of the Piedmont province of the Appalachian orogen on the west to relatively undeformed, slightly dipping, Mesozoic and Cenozoic sediments of the Coastal Plain on the east. Originally, many Cenozoic marine sequences extended westward onto the Piedmont beyond the location of the present-day Fall Line, but large-scale Cenozoic erosion removed much or all of the shallow-marine facies deposited during the higher sea level stands (Gibson *et al.,* 1992).

We will examine briefly the relatively poorly known record of Paleocene-Eocene boundary sedimentation in 3 southern Atlantic Coastal Plain embayments and then concentrate on the Salisbury embayment to the north in which New Jersey contains the only known U.S. section with continuous marine sedimentation across the Paleocene-Eocene boundary.

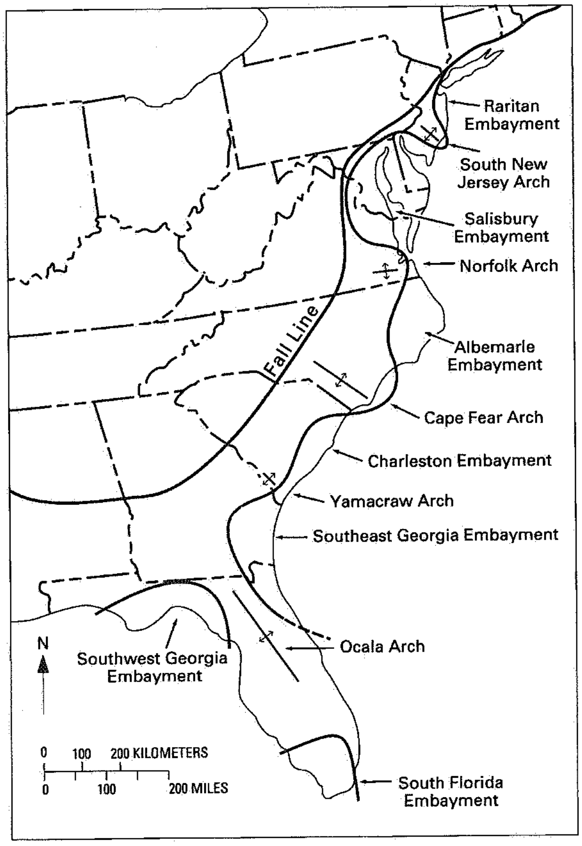

Figure 7. Map showing major embayments and arches of the U.S. Atlantic Coastal Plain.

5.1. Southern Atlantic Coastal Plain

Significantly less is known about the stratigraphy of the Paleocene-Eocene boundary interval in the Albemarle, Charleston, and Southeast Georgia embayments (Fig. 7) than is known from the eastern Gulf of Mexico Coastal Plain and the Salisbury embayment. Paleocene strata do crop out in these 3 southern embayments, but lower Eocene marine strata are known only from the subsurface of the 3 embayments, and there are few continuously cored drillholes that penetrate this interval.

During the late Paleocene and early Eocene there was a facies change southward from dominantly terrigenous sediments in North Carolina to carbonate deposits in Florida (Gohn, 1988). Terrigenous clastic sedimentation dominated the Paleocene strata in North Carolina and South Carolina, whereas Georgia contained the transition to the carbonate lithology characteristic of the Florida deposits. Few terrigenous sediments were transported into Florida because of its distance from the source areas. The transition zone between the carbonate province and terrigenous province shifted northward during the early Eocene, and carbonate sedimentation extended as far north as SW South Carolina (Gohn, 1988). There are no detailed biostratigraphic data from these upper Paleocene and lower Eocene carbonate deposits because calcareous microfossils commonly are sparse and recrystallized. Gohn (1988) discussed stratigraphic uncertainties of the Paleocene and Eocene deposits in his summary of the early Cenozoic history of the North Carolina to Florida coastal plain.

The carbonate and evaporite sections in Florida have yielded few calcareous planktonic microfossils. The Oldsmar Formation (Fig. 8), composed of dolomite, limestone, and sparse evaporite beds, is considered to be late Paleocene and early Eocene in age (Gohn, 1988), but a more refined age is not yet available.

In SE Georgia, the Davis-Hopkins corehole (Fig. 2) penetrated carbonate sediments belonging to Zones NP 9 and NP 10 (Bybell, unpubl. data), but a more precise placement within these 2 zones was not possible from the limited assemblages that are present. In Geor-

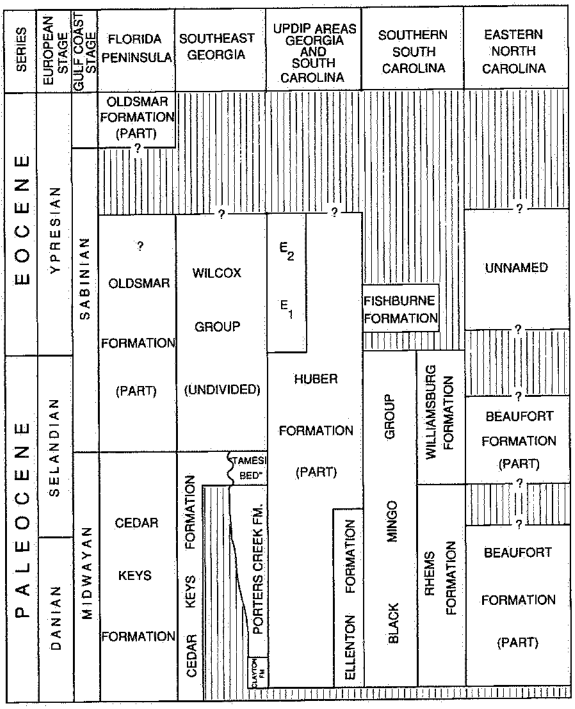

Figure 8. Correlation chart of Paleocene and lower Eocene formations of the southern Atlantic Coastal Plain (from Gohn, 1988).

gia and South Carolina, sparingly fossiliferous sand, kaolinite beds, and lignitic clays of the Huber Formation (Fig. 8), considered by Gohn (1988) to represent fluvial and delta plain deposits, occur in upbasin areas. There is no detailed biostratigraphic analysis of these commonly highly weathered strata.

The biostratigraphic succession is som-ewhat better understood in southern South Carolina. The Williamsburg Formation (Fig. 8) consists of clay, sand, bioclastic limestone, and sequences of thinly interbedded sand and dark clay (Van Nieuwenhuise & Colquhoun, 1982 ; Gohn, 1988). In the Williamsburg Formation, Gohn (1988) delineated a fine-grained, fossiliferous unit that he considered to have been deposited in prodelta-shelf environments ; it is overlain by an upper unit of well-sorted sand, shell beds, and thinly interbedded sand and dark clay that he considered to have been deposited in nearshore and marginal-marine environments. In sequence stratigraphic concepts, the Williamsburg appears to consist of a transgressive systems tract overlain by a highstand regressive tract. Whether the Williamsburg sequence corresponds to any of the upper Paleocene eastern Gulf Coast sequences recognized by Mancini & Tew (1991) is unknown at the present time because of the lack of precise biostratigraphic control in South Carolina. In the Clubhouse Crossroads #1 corehole (Fig. 2), the Williamsburg was assigned to planktonic foraminiferal Zones P 4-5 and calcareous nannofossil Zones NP 5 to NP 9 (Gohn, 1988) ; strata from the upper part of the formation contain the calcareous nannofossil *Discoaster multiradiatus* without the presence of the genus *Tribra-chiatus* and are presumed to belong in Zone NP 9.

A glauconitic, clayey, finely crystalline limestone unit, the Fishburne Formation, which is known only from the subsurface, disconformably overlies megafossiliferous, quartzose limestone at the top of the Paleocene Williamsburg Formation in South Carolina (Gohn *et al.,* 1983). A limited calcareous nannofossil assemblage indicated an early Eocene age for this unit. The presence of the dinoflagellate genus *Wilsonidium* at the base of the Fishburne type section in the Clubhouse Crossroads #1 corehole (Gohn *et al.,* 1983) allows a more precise stratigraphic placement. Edwards (1989) reported the FAD of *Wilsonidium* some distance above the base of lower Eocene deposits in Virginia. In Alabama, the FAD of *Wilsonidium* occurs in the fifth cycle of the Bashi (L.E. Edwards, personal commun., 1992) along with the calcareous nannofossils *Tribrachiatus contortus* and *T. orthostylus* (Fig. 5) ; this co-occurrence places the FAD of *Wilsonidium* in the upper half of Zone NP 10. Thus, the lower part of the Fishburne Formation at the type section belongs in upper Zone NP 10 or higher, and the hiatus between the Williamsburg and Fishburne Formations represents at least the lower half of Zone NP 10.

The Albemarle embayment in North Caro-lina contains dominantly terrigenous Paleocene and lower Eocene deposits ; the major northward movement of carbonate lithologies did not reach North Carolina until the middle Eocene. There are few stratigraphic data available concerning upper Paleocene and lower Eocene strata in this area because few outcrops and coreholes contain these deposits. Recently, the U.S. Geological Survey drilled a continuously cored hole at Valhalla, North Carolina (Fig. 3), on the northern flank of the Albemarle embayment that did recover Paleocene and Eocene strata. In the Valhalla corehole, glauconitic sand of the Beaufort Formation of early late Paleocene age (Zone NP 5) is disconformably overlain by a 3-m-thick shelly carbonate unit of late Paleocene age, which may correlate with the limestone unit present at the top of the Williamsburg Formation in South Carolina. Limited calcareous nannofossils in this unnamed carbonate sequence indicate that its age lies within the Zone NP 5 to middle of Zone NP 9 interval (Bybell, unpubl. data). Disconformably overlying this carbonate bed is a glauconitic clayey sand that contains *Discoaster lodoensis*, but not *Tribrachiatus orthostylus* or *Discoaster sublodoensis*, and this unnamed sand unit is placed in Zone NP 13. The Valhalla corehole demonstrates that at least in this part of the Albemarle embayment there is a significant disconformity across the Paleocene-Eocene boundary that represents the uppermost part of Zone NP 9 and all of Zones NP 10 to NP 12.

5.2. North-central Atlantic Coastal Plain

The most complete marine record of Paleocene-Eocene boundary sediments in outcrop and shallow subsurface exposures in the eastern U.S. occurs in the Salisbury embayment (Fig. 3). The Salisbury embayment is north of the Paleogene carbonate deposition region of the Atlantic Coastal Plain. Strata in this embayment consist dominantly of highly bioturbated, glauconitic, clayey, silty, very fine to fine-grained sand units that were deposited at relatively low sedimentary rates into a shallow, low-energy marine basin. Glauconite is abundant in most beds and in some intervals exceeds 50 % of the sand-sized fraction.

In the Salisbury embayment, the Paleogene sedimentary rates are the lowest for the entire Mesozoic and Cenozoic in this region (Poag & Sevon, 1989). The low sedimentary rates that are typical of the late Paleocene and early Eocene are exemplified in the Potomac River outcrops in Virginia and Maryland (Fig. 3) ; there, 3 formations with an outcrop thickness totaling 60 m repres-ent all calcareous nannofossil zones from NP 5 to NP 13. The thinness of the units in the central and northern Atlantic Coastal Plain makes them more susceptible to subsequent partial or complete removal by erosion than the considerably thicker coeval Gulf Coastal Plain sequences. In the Potomac River outcrop belt in Virginia, some thin, glauconitic, quartz sand beds with a thickness of 0.5-2 m or less may represent an entire calcareous nannofossil zone (Gibson & Bybell, 1991). We consider that these thin units are the lower part of transgressive units that remain after subsequent erosion removed the upper prograding part of the transgressive-regressive cycle and cut down into the transgressive portion. In this region of low sedimentation rates, the eroded portion of a sequence may have been fairly thin. The preserved beds also are thin, and their presence in the basal part of a cycle as indurated shelly sand, indicates that their resistance to erosion probably was a major factor in their not being removed.

Marine deposition persisted later into Zone NP 9 in the southern and central parts of the Salisbury embayment, as seen in the present-day outcrop belt of uppermost Paleocene strata, than it did in the eastern Gulf Coast (Frederiksen *et al.,* 1982). The U.S. Atlantic margin never received the large detrital influx that in the Gulf Coast resulted in the thick prograding wedge of late Zone NP 9 sediments of the Tuscahoma Formation.

When originally deposited, upper Pale-ocene and lower Eocene sedimentary facies were similar in the northern and southern parts of the Salisbury embayment. However, subsequent differences in uplift and erosional history between the northern and southern parts of the embayment resulted in different facies now being present in the outcrop and shallow subsurface sections of the 2 areas. During the Miocene, New Jersey and the northern Appalachian area underwent relatively greater uplift than did areas to the south (Gibson, 1970 ; Owens, 1970 ; Poag & Sevon, 1989). The uplift caused increased erosion of the more elevated western areas of the Piedmont and resulted in the removal of the upbasin shallower water facies of Paleogene units from those areas. This uplift also raised to the surface and/or shallow subsurface the downbasin, deeper water facies equivalents of the eroded strata, as well as strata belonging to depositional sequences that were formerly present only in the deeper subsurface. Because there was less uplift in the central and southern parts of the Salisbury embayment, downbasin facies and units in this region are likely to be in the deeper subsurface ; they are largely unknown because they are infrequently, if at all, penetrated by continuous coreholes. Thus, as Poore & Bybell (1988) demonstrated, a much more complete record of Eocene sedimentation is present in the shallow subsurface of the New Jersey Coastal Plain than is presently known from the subsurface of Maryland and Virginia.

The Miocene uplift caused the differences that we now see in the record of uppermost Paleocene and lower Eocene sedimentation in the northern versus the southern parts of the Salisbury embayment. The Aquia Formation and the Marlboro Clay comprise the upper Paleocene strata in the southern and central part of the embayment, and the only deposits of these formations that are found in the present-day outcrop and shallow subsurface belt were deposited in shallow-marine environments (Gibson, Andrews *et al.,* 1980).

In New Jersey (northern Salisbury embayment), upper Paleocene and lower Eocene shallow-water strata were located on the western, more uplifted parts of the Piedmont, and subsequent erosion has removed these strata (Gibson *et al.,* 1992). The uplift, however, also caused deeper water, middle to outer neritic facies of these units, which normally would occur at moderate depths in the subsurface of the coastal plain, to be brought into the shallow subsurface near the present western margin of the New Jersey Coastal Plain. These deeper water sections were penetrated in the GL and Clayton drillholes (Fig. 3), and they contain continuous sedimentation across the Zone NP 9-Zone NP 10 boundary because they were deposited in water deep enough to be unaffected by the lowered sea levels that were present at or near the end of Zone NP 9.

5.2.1. Upper Paleocene Lithostratigraphy

The shelly, variably clayey, glauconitic quartz sands of the Aquia Formation in Maryland and Virginia (Fig. 3) reflect the low sedimentation rates in shallow-marine environments. The medium- to dark-olive-gray, very fine to medium-grained, massive sand usually is abundantly glauconitic, ranging from a low of 5-10 % to well over 50 % glauconite in the sand fraction. The lower part of the formation has poorer sorting and more clay than the sandier, coarser grained, better sorted upper part. Mollusk shells are abundant in many beds, either randomly scattered or in lenses, or as thick, laterally persistent lag deposits. Although a diversified molluscan fauna is present in the outcrops (Ward, 1985), some Aquia intervals are heavily dominated by lenses of oysters or *Turritella.* Several carbonate-cemented, shelly sand beds, 0.2-0.7 m thick, are present in many areas and form prominent marker beds in outcrop exposures.

Late Paleocene calcareous nannofossil Zones NP 5 through NP 9 are present in approximately 25 m of Aquia strata along the Potomac River (Gibson & Bybell, 1984 ; Bybell & Gibson, 1991). Some of these zones, however, are represented by only 1-2 m of sediment (Gibson & Bybell, 1991), and it is presumed that only a small portion of the total time interval represented by some zones is recorded in these strata. The thinness of the shelly, glauconitic, quartz sand strata, the presence of numerous scour surfaces and more obvious disconformities, and the rapid biostratigraphic succession suggest that relatively incomplete portions of transgressive units are preserved here. In the Salisbury embayment, there is a general absence of the thick carbonaceous, interbedded sand, silt, and clay deposits that compose the prograding upper part of sequences, the clearly recognizable highstand systems tract, that are so prevalent in this time interval in the eastern Gulf Coastal Plain. The Aquia Formation thickens downbasin to the east ; it is 35 m thick in the Oak Grove corehole, 46 m thick in the Solomons Island corehole (Fig. 3), and 73 m thick in the subsurface to the NE of Solomons Island (Hansen, 1974).

Aquia strata belonging to Zone NP 8 are very thin in the Potomac River outcrop sections (less than 1 m) (Gibson & Bybell, 1991), and they represent only a partial record of both this zone and its associated transgressive sequence. Farther downbasin, strata of Zone NP 8 thicken to 8 m in the Oak Grove corehole and to 13 m in the Solomons Island corehole (Gibson, Andrews *et al.,* 1980). In contrast, the prograding, very shallow water, well-sorted sand beds of Zone NP 9 of the Aquia are 12-15 m thick in the Potomac River outcrop sections, but only about 5 m thick in the Oak Grove corehole to the east.

Most of the outcropping Aquia strata were deposited in inner neritic environments (Nogan, 1964 ; Beauchamp, 1984 ; Gibson, unpubl. data). Inner neritic environments also are indicated for many of the Aquia sediments in the downbasin coreholes (Gibson, Andrews *et al.,* 1980 ; Poag, 1989). A few intervals, particularly those in the lower and middle parts of the Aquia in these coreholes, contain deposits suggestive of inner middle to middle neritic environments. These paleobathymetric interpretations were based upon the species and generic composition and diversity of the benthonic foraminiferal assemblage and upon the planktonic foraminiferal species diversity and the percentage of planktonic specimens in the entire assemblage. An examination of the *tau* values (*tau* = number of benthonic species x planktonic percentage ; the values generally increase with increasing water depth, see Gibson, 1988) for the foraminiferal assemblages in the Oak Grove corehole (Fig. 9) suggests a similar inner to middle neritic paleobathymetry. The *tau* values in most Aquia samples are less than 300 and suggest inner neritic environments with water depths less than 40-50 m. The samples in Zone NP 6/7 and lower Zone NP 8 have *tau* values of 300-900, along with increased proportions of *Tappanina selmensis* and *Trifarina wilcoxensis*, and suggest deposition in inner middle neritic environments with water depths of 50-100 m.

The uppermost glauconitic sand beds of the Aquia Formation, which are placed in Zone NP 9 (Gibson & Bybell, 1984 ; Bybell & Gibson, 1991), were deposited under higher energy conditions in very shallow inner neritic environments in the outcrop belt (Nogan, 1964 ; Beauchamp, 1984) and in shallow inner neritic environments in the more downbasin coreholes (Gibson, Andrews *et al.,* 1980 ; Poag, 1989). These relatively well-sorted sand beds are at least 12 m thick and comprise the majority of the total formation in the type area (Gibson & Bybell, 1991) ; this thickness, along with an upward increase in very shallow water foraminiferal species, suggests that these upper beds represent prograding deposits into shoaling, very shallow marine environments. These sands probably are part of the offshore sand bank complex that Hansen (1974) proposed for the upper part of the Aquia.

Two depositional cycles are recognized in the upper part of the Aquia Formation in outcrops in the type area along the Potomac River in Virginia (Fig. 3, #5). The lower one includes a shelly, clayey, glauconitic sand at the base and a non-megafossiliferous sand at the top (Zones NP 8 and NP 9). This cycle is separated by a strongly burrowed surface from the overlying second cycle of relatively thick fossiliferous sand. The precise age relationships between these cycles and those present in the Tuscahoma Formation of Alabama is unknown at the present time.

There is a disconformity between the Aquia and Marlboro in some outcrop sections (Ward, 1985 ; Gibson *et al.,* 1991 ; Gibson & Bybell, 1991) ; however, in other sections, such as in the Oak Grove corehole, there is a rapid upward gradation over an interval of about 0.1 m that exhibits interfingering of glauconitic sand of the Aquia with clay of the Marlboro (Reinhardt *et al.,* 1980). Glaser (1971) proposed that the surface between the 2 units was disconformable in upbasin areas and that this surface became conformable with a thin transition zone in more downbasin areas.

The Marlboro Clay is a relatively thin unit that has a widespread distribution throughout southern Maryland and Virginia ; it extends for nearly 200 km from north to south and for 72 km from west to east (Glaser, 1971). The Marlboro reaches a maximum thickness of 9.1 m (Glaser, 1971) but commonly varies between 0.1-6 m. Rapid lateral variations in thickness result from varying amounts of erosional truncation at the disconformity that is always present at the top of the Marlboro. The Marlboro is a massive to finely laminated clay with considerable silt content. The beds commonly are mottled and burrowed ; the degree of bioturbation is variable within the mottled intervals, but usually it is sufficient to obliterate any primary deposi-tional structures. Silt is present in thin laminations as well as in thicker silt beds (Glaser, 1971). The laminations, usually 1 mm thick, are defined either by slight color changes or by varying amounts of silt ; ripple cross laminations may be present, particularly in the siltier intervals (Glaser, 1971 ; Reinhardt *et al.,* 1980). Lignite grains may be present in the silt beds (Glaser, 1971). The clay beds sometimes contain small, 0.5 cm-sized pods of fine-grained glauconitic sand. The clay mineralogy changes from a smectite and illite dominated suite in the Aquia Formation to a kaolinite dominated suite in the overlying Marlboro (Reinhardt *et al.,* 1980 ; McCartan, 1989). The Marlboro is dominantly pink to red-brown in color in the outcrops and shallow coreholes ; it commonly contains a considerably lesser amount of gray clay in the lower and upper parts of the formation. A burrowed, disconformable surface between the Marlboro and the overlying Nanje-moy Formation is present in all observed sections (Gibson, Andrews *et al.,* 1980 ; Ward, 1985 ; Mixon *et al.,* 1989).

Most previous studies postulated restricted-marine to brackish-water depositional environments for the Marlboro Clay, based largely upon the presence of low-diversity assemblages of agglutinated foraminifers (Nogan, 1964 ; Gibson, Andrews *et al.,* 1980) and dinoflagellate assemblages in which *Senegalinium* ? *dilwynense* is the only abundant species (Edwards *et al.,* 1984). Only a few Marlboro samples from outcrops and shallow subsurface coreholes contain foraminifers, and the assemblages from these localities usually contain only agglutinated taxa (Nogan, 1964 ; Gibson, Andrews *et al.,* 1980). Occasionally a few poorly preserved gastropods and pelecypods are present.

Our recent investigations yielded calcare-ous foraminiferal assemblages from the Marlboro Clay in the Waldorf, Maryland, and Putney Mill, Virginia, coreholes (Fig. 3). These coreholes are near the northern and southern ends of the Marlboro depositional area outlined by Glaser (1971). The foraminiferal assemblages from these 2 coreholes suggest that the entire Marlboro was deposited in shallow-marine environments. However, the unusual composition of the assemblages and the rarity of preserved calcareous foraminifers in the cores suggest that abnormal environmental conditions were present in inner to middle neritic depths during deposition of the Marlboro. The sporadic occurrence of calcareous assemblages in the Marlboro suggests that although these sediments were deposited in neritic depths, either some unknown environmental conditions prohibited them from living there during much of Marlboro time or they were subsequently removed by dissolution. The few agglutinated-only assemblages could reflect environmental conditions that were suitable only for their existence, or they could be the residuum from the dissolution of mixed calcareous-agglutinated assemblages. The absence of agglutinated specimens in the Marlboro calcareous assemblages suggests the presence of the former condition.

Calcareous foraminiferal assemblages were obtained from the Putney Mill corehole from 2 samples at a 1-m spacing in the lower part of the Marlboro ; samples examined from other parts of the Marlboro in this corehole were barren of foraminifers. Here, the lower Marlboro here is a massive, red-brown, silty, micaceous clay containing some 1 mm laminations and pods of glauconitic fine sand. *Pulsiphonina prima* is the dominant species in both assemblages and composes 76 and 81 % respectively of the total benthonic specimens. Fifteen benthonic species occur in an aliquot of the lower sample, and 17 species occur in the upper sample. Other benthonic taxa occurring in small proportions in both assemblages are *Eponides mexicanus* and *Gyroidinoides* sp. *Cibicides alleni* occurs as a small component only in the lower sample. *Pyramidina virginiana* is a much larger component of the upper sample than of the lower sample. The lower sample contains 2.6 % planktonics, and the upper one contains 1.6 % ; the specimens in both samples consist of immature specimens of *Sub-botina.*

The species composition of the moderately diverse benthonic foraminiferal assemblages, the presence of small proportions of planktonic specimens, and the presence of a low-diversity calcareous nannofossil assemblage in these samples suggest that they were deposited in shallow-marine environments that were either completely open to oceanic circulation or at most had only partially restricted access. The unusually high proportion of *Pulsiphonina* in the benthonic foraminiferal assemblages, along with the lesser proportion of *Pyramidina virginiana*, suggests that abnormal environmental conditions, possibly those of high productivity and/or lowered oxygen levels, were present in these shallow-marine environments near the southern margin of the Marlboro depositional area. *Pulsiphonina* also occurs in the deeper water, low-oxygen environments that are present in the lower part of the Manasquan Formation in New Jersey at this time, but it occurs there in considerably lower abundances. The lower Marlboro Clay in the Putney Mill corehole is red ; this color suggests that although lower than normal oxygen conditions may have been present, oxygen levels were not low enough to cause reduction of the iron in the sediment.

The upper part of the Marlboro Clay in the Waldorf corehole also has unusually low foraminiferal diversity, combined with the dominance of a species that normally composes a relatively small part of the benthic assemblage. Only 1 Marlboro sample in this corehole contains foraminifers, and it has a low diversity assemblage (11 species). *Anomalinoides acutus*, which usually occurs in small to moderate abundances in middle neritic environments, composes 58.6 % of the assemblage. Planktonic specimens compose 7.6 % of the assemblage. The assemblage suggests abnormal inner to middle neritic environments for the Marlboro here.

Combined data from the eastern Gulf Coast (Frederiksen, 1992) and the middle Atlantic Coastal Plain (Gibson, Andrews *et al.,* 1980) indicate that a prominent late Paleocene extinction event took place in the terrestrial flora throughout eastern North America. This extinction of approximately 13 taxa occurred within the relatively short time period represented by the deposition of the Marlboro Clay in upper-most Zone NP 9. The Marlboro Clay also contains the first occurrence in the middle Atlantic Coastal Plain of 14 taxa (Gibson, Andrews *et al.,* 1980). Many of these 14 taxa occur in older strata in the Gulf Coastal Plain (N.O. Frederiksen, personal commun., 1992) ; there may have been a significant migration of Gulf Coast floras into this more northerly area during the deposition of the Marlboro because of the warming terrestrial climate that occurred at this time (Frederiksen, 1979 ; REA *et al.,* 1990). Alternatively, the absence of these species in pre-Marlboro sediments in the Atlantic Coastal Plain simply may be the result of there being no samples from the appropriate paleoenvironments. A large number of fern spores in Marlboro deposits in the Oak Grove corehole suggests moist conditions in the sediment source area. The Marlboro is the only Cenozoic unit in the Oak Grove corehole that contains reworked palynomorphs, those being of Paleozoic and Late Cretaceous age, and their occurrence also may indicate more intense erosion in the sediment source areas at this time.

The thin, widespread nature of the Marlboro Clay led Mixon *et al.* (1989) to suggest that this unit was deposited in restricted, shallow shelf waters or, possibly, in a large barred embayment rather than being deposited in an estuary. The calcareous, marine foraminiferal assemblages that sporadically occur in the Marlboro suggest that this unit was deposited in shallow-marine environments, albeit unusual ones ; the planktonic component of the assemblages, however, suggests that completely open marine environments probably were present. These occurrences suggest that Virginia and southern Maryland were part of a low-energy mud-dominated coastal area during the deposition of the Marlboro Clay and that mud accumulated here in shallowto middle-shelf environments within the Salisbury embayment. This mud-dominated shelf extended into New Jersey where the clay of the lower part of the Manasquan Formation was deposited in middle neritic environments at this time (see 5.3.3). The Valhalla, North Carolina corehole (Fig. 3) did not penetrate any clay unit in this stratigraphic position, but the large disconformity present there may include latest Paleocene as well as early Eocene time.

5.2.2. Upper Paleocene Biostratigraphy

The upper strata of the Aquia Formation were deposited in shallow-marine environments and they are more accurately dated with calcareous nannofossils than with the sparse, low-diversity, planktonic foraminiferal assemblages. The Aquia strata placed in Zone NP 8 contain *Heliolithus riedelii* ; the strata placed in Zone NP 9 contain *Discoaster multiradiatus*, but no *Tribrachiatus bramlettei.*

The upper part of the Aquia contains an influx of dinoflagellate species, including *Kallosphaeridium brevibarbatum* and *Impagidinium* sp. (Edwards, 1989). *Apectodinium homomorphum*, which first appears near the base of the upper part of the Aquia, approx-imates the base of Zone NP 9, and this species becomes increasingly abundant toward the top of the formation (Edwards *et al.,* 1984).

Frederiksen *et al.* (1982) suggested that the floral change within the Marlboro Clay may coincide with the Paleocene-Eocene boundary ; at that time no calcareous microfossil data were available for the Marlboro. Subsequently, calcareous nannofossils indicative of Zone NP 9 (Bybell, unpubl. data) were found near the top of the Marlboro section in the Waldorf, Maryland corehole (Fig. 3), which suggests that, at least there, the entire Marlboro may be latest Paleocene in age. *Tribrachiatus bramlettei* (occurrence restricted to Zone NP 10) is absent from the Marlboro. This species commonly is found in basal Nanjemoy Formation samples. However, in recently examined samples from the Loretto, Virginia, corehole (Fig. 3), *T. bramlettei* is absent from the lower several feet of the Nanjemoy Formation. This species also is absent from the lowest Nanjemoy sample in the Oak Grove corehole. These data suggest that in some areas the basal Nanjemoy may belong in uppermost Zone NP 9. If so, the underlying Marlboro is confined to Zone NP 9 and does not contain the Zone NP 9-NP 10 boundary.

5.2.3. Lower Eocene Lithostratigraphy

The Nanjemoy Formation is a poorly sorted, very fine to fine-grained glauconitic quartz sand that contains considerable, but varying, amounts of clay and silt. In the Oak Grove corehole, the amount of clay varies from 15-80 %, and glauconite composes as much as 65 % of the sand fraction in the Nanjemoy (Reinhardt *et al.,* 1980). Lignitic debris and mica also occur in the formation. The upper part of the formation generally is less clayey and may contain scattered coarse sand to fine gravel-sized quartz. The unit is generally massively bedded, and abundant bioturbation obscures primary bedding structures. Kaolinite is abundant in the lowermost part of the Nanjemoy (Reinhardt *et al.,* 1980), but illite and smectite dominate the clay mineral suite in all of the middle and much of the upper part of the formation (McCartan, 1989). Molluscan shells of various taxa are commonly scattered through

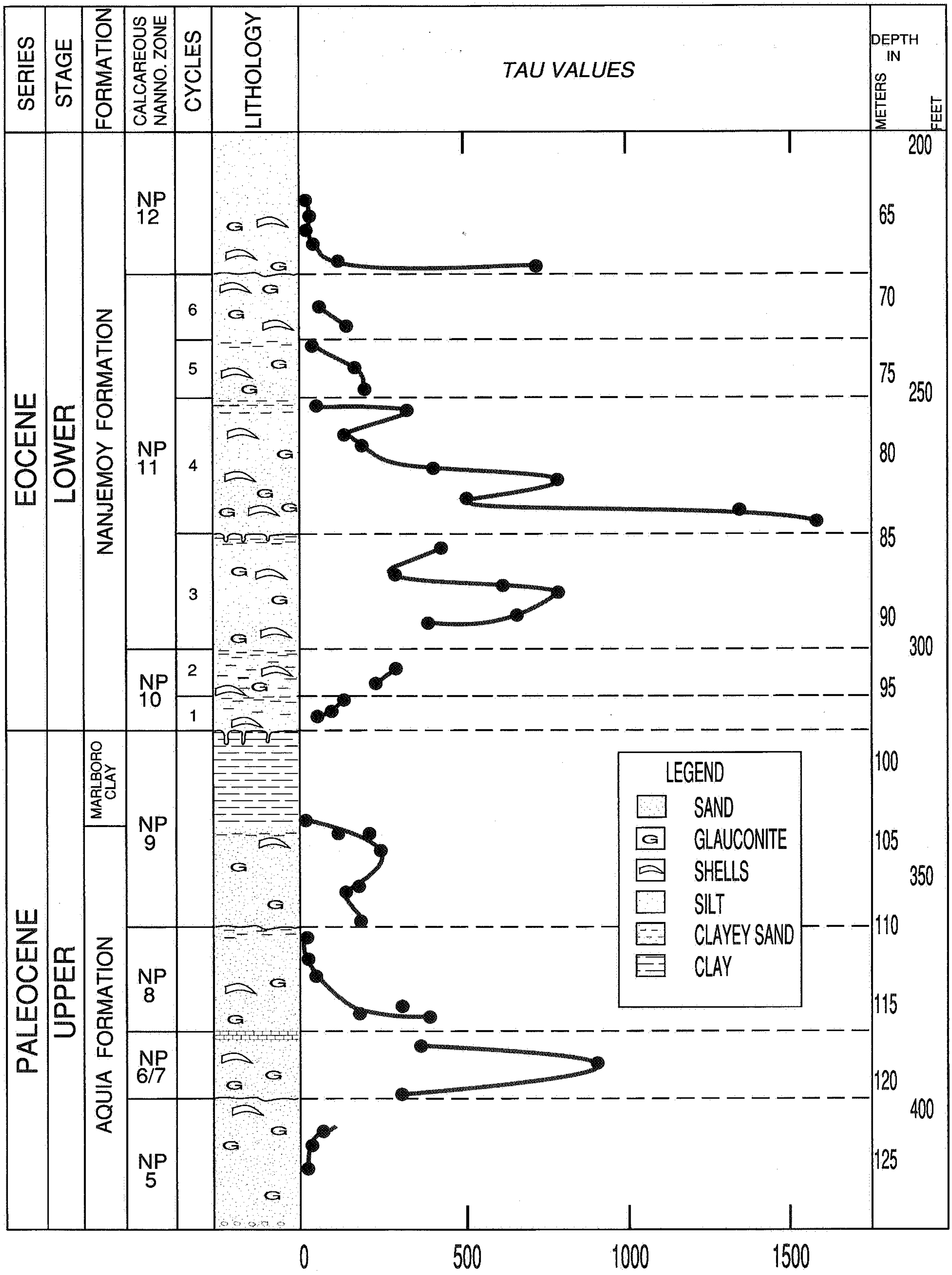

Figure 9. Lithologic section and *tau* values (planktonic % X number of benthonic foraminiferal species) of upper Paleocene and lower Eocene strata in the Oak Grove, Virginia, corehole ; note deepest water deposition in cycle 4 during Zone NP 11.

the formation ; layers of *Venericardia* are common in the lower part of the Nanjemoy.

The thickness of the Nanjemoy was esti-mated at 26-32 m from outcrops along the Potomac River (Clark & Martin, 1901). There are only a few discontinuous, relatively thin outcrops of the lower part of the Nanjemoy along the Potomac, however, and it is difficult to study this part of the formation from surface exposures. The coreholes, however, contain continuous sections of the lower beds. The Nanjemoy is 37.4 m thick in the Oak Grove corehole (Reinhardt *et al.,* 1980) and increases to a thickness of 47.8 m in the Solomons Island corehole (Gibson, unpubl. data). Reinhardt *et al.* (1980) recognized 3 units in the Nanjemoy Formation in the Oak Grove corehole : a basal clayey sand that is overlain by 2 sequences with thicknesses of 10-15 m, each of which coarsens upward from sandy clay to medium sand. These 3 units belong in Zones NP 10 to NP 12 (Gibson, Andrews *et al.,* 1980). Mixon *et al.* (1989) recognized 4 fining upward sequences in the lower and middle parts of the Nanjemoy in the more downbasin Haynesville, Virginia, corehole (Fig. 3) and presumably 1 sequence in the upper part of the Nanjemoy. Although calcareous nannofossil data are not available from the Haynesville corehole, dinoflagellate assemblages (Edwards, 1989) suggest that the lower 4 sequences belong in Zones NP 10 to uppermost NP 11 or lowermost NP 12. Ward (1985) also divided the lower part of the Nanjemoy into 4 beds, which he considered to represent possibly 3 depositional events. Mixon *et al.* (1989) discussed the possible correlation of Ward's units with those delineated in the Haynesville corehole. Burrowed and scoured surfaces are present both in outcrop and corehole sections of the Nanjemoy, but the biostratigraphic data needed to provide detailed placements of these possible sequence boundaries within Zones NP 10 to NP 13 is still lacking for many sections.

Our restudy of the Oak Grove corehole indicates that 6 fining upward depositional cycles can be recognized in strata of Zone NP 10 to NP 11 age (Fig. 9). Here, the physical characteristics that define the cycles are much more subtle than those present in the Bashi Formation in Alabama. The lower part of the Oak Grove cycles is composed of glauconitic sand, while the top of the cycles is marked by an increase in the amount of clay, a decrease in glauconite, and by the sparseness or absence of calcareous microfossils. Although illite and smectite are the dominant clay minerals throughout most of each cycle, the uppermost clayey part of a cycle contains a significant amount of kaolinite (Gibson, unpubl. data) ; this suggests that the highly kaolinitic Marlboro Clay was being eroded during this time of progradation and was redeposited into the uppermost, shallowest environments of the Nanjemoy cycles. Two cycles of Zone NP 10 age and 4 cycles of Zone NP 11 age are present in the Oak Grove core. Both the second cycle of the Nanjemoy Formation in the Oak Grove corehole and the fifth cycle of the Bashi Formation in Alabama are placed in the upper half of Zone NP 10. Thus, there are 4 cycles of lower Zone NP 10 age preserved in Alabama, but only 1 in Virginia. However, there are 4 cycles of Zone NP 11 age preserved in Virginia compared with only 2 of this age in Alabama. Additional work is needed to determine whether these regional differences are the result of erosion or original deposition. The present thickness of Zone NP 10 strata differs between the two areas. Zone NP 10 strata of the Bashi and Hatchetigbee Formations are 30 m thick in Coffee County, Alabama, whereas Zone NP 10 strata of the Nanjemoy Formation are only 6 m thick in Virginia and Maryland. Significant truncation of the Nanjemoy Formation after deposition controlled how much of the upper part of the Nanjemoy is preserved in different areas of the Salisbury embayment ; for example, the entire upper portion of the Nanjemoy is missing in upbasin areas. Thus, the upbasin Waldorf corehole (Fig. 3) contains strata only of Zones NP 10 and NP 11. The somewhat farther downbasin Oak Grove corehole contains deposits of Zones NP 10, NP 11, and NP 12, and the more downbasin Solomons Island corehole contains Nanjemoy deposits of Zones NP 10, NP 11, NP 12, and NP 13. As the proposed sea levels for this area were higher during Zones NP 12 and NP 13 than those during Zone NP 10 (Fig. 9), the younger portion of the Nanjemoy originally should have been more widespread than the Zone NP 10 strata. In addition to regional truncation, some local variations in both the presence and the thickness of calcareous nannofossil zones that are seen both in coreholes and in outcrops may be the result of penecontemporaneous faulting (Mixon & Powars, 1984).

Inner neritic depositional environments of 30-60 m depth or slightly less were proposed for most of the Nanjemoy Formation (Gibson, Andrews *et al.,* 1980 ; Poag, 1989). Low *tau* values (Fig. 9) also indicate inner neritic environments for these beds. Sediments from Zone NP 11 in the Oak Grove, Putney Mill, and Solomons Island coreholes represent the deepest water deposition found in the Nanjemoy, that is middle neritic depths of 100 m or slightly deeper in the Oak Grove and Solomons Islands coreholes (Gibson, Andrews *et al.,* 1980 ; Gibson, unpubl. data). This determination was based upon *tau* values (Fig. 9), which also indicate that middle neritic depths were present in cycle 4 in Zone NP 11 deposits and that shallowing occurred in the uppermost part of this zone in the Oak Grove corehole. Poag (1989) proposed that the deepest water deposition in the Nanjemoy from the Haynesville corehole occurred at depths of 100-150 m, based on a sample that he placed in the uppermost part of planktonic foraminiferal Zone P6b. Although a calcareous nannofossil zonation is not available for the Haynesville corehole, Poag's stratigraphic placement in upper Zone P6a of these deeper water deposits correlates with the Zone NP 11 placement of the deeper water strata in other Virginia and Maryland coreholes.

Numerous benthonic genera, which occur in the clayey sand beds of the lower part of the Nanjemoy in Virginia and Maryland, suggest that high-nutrient and/or low-oxygen bottom conditions were present in these earliest Eocene shallow-water environments. These genera include *Bolivina, Epistominella, Florilus, Globobu-limina, Lenticulina, Loxostomum, Pseudouvigerina, Pulsiphonina, Pyramidina*, and *Turrilina* (Gibson, Andrews *et al.,* 1980 ; Poag, 1989 ; Gibson, unpubl. data). These genera are most abundant in Zone NP 10 strata ; smaller proportions of some of these genera also occur in some middle and upper intervals of the Nanjemoy. In addition to the above genera, Zone NP 10 beds that were deposited in upbasin, shallower water localities, including the outcrops along the Potomac River, contain high percentages of *Buliminella elegantissima.* In contrast, Zone NP 10 samples from deeper water environments in more downbasin coreholes contain moderate abundances of *Tappanina selmensis.*

5.2.4 Lower Eocene Biostratigraphy

The Nanjemoy Formation was placed in Zones NP 10 to NP 13 (Bybell & Gibson, 1991) ; as mentioned earlier, the basal strata of the Nanjemoy in 2 coreholes may belong in Zone NP 9. The first appearance datum (FAD) of *Tribrachiatus bramlettei* marks the base of Zone NP 10. The last appearance datum (LAD) of *Tribrachiatus contortus* is used to mark the top of Zone NP 10, and the FAD of *Discoaster lodoensis* marks the base of Zone NP 12. The LAD of *Tribrachiatus orthostylus* is the marker for the top of Zone NP 12. Sediments located above the LAD of *T. orthostylus* and which do not contain *Discoaster sublodoensis* are placed in Zone NP 13.

Dinoflagellate studies of the Nanjemoy include Goodman (1979 ; 1984), Gibson, Andrews *et al.* (1980), Edwards *et al.* (1984), and Edwards (1989). These studies discussed possible correlations of the Nanjemoy dinoflagellate assemblages, particularly those from the middle and upper parts of the formation, with the northwestern European dinoflagellate zonations. Edwards *et al.* (1984) listed a succession of acritarch and dinocyst first occurrences in the Nanjemoy that have probable age significance ; in ascending order they are *Ascostomocystis hydria, Wilsonidium tabulatum, Wetzeliella hampdenensis, Biconidinium longissimum, Emmetrocysta* sp. of Edwards *et al.* (1984), *Homotryblium tasmaniense, Achilleodinium biformoides, Wetzeliella varielongituda/samlandia, Kisselovia coleothrypta, Homotryblium caliculum, H. tenuispinosum/pallidum, Hafniasphaera goodmanii, Spinidinium* sp. of Goodman (1979), and *Wetzeliella* sp. of Goodman (1979). Some of these first occurrences, particularly those in the middle and upper parts of the Nanjemoy, may prove valuable in correlation with northwestern European deposits.

5.3. Northern Atlantic Coastal Plain

As discussed earlier, the northern part of the Salisbury embayment was uplifted during the late Oligocene to middle Miocene. This uplift brought both lower Paleocene Zone NP 3 and NP 4 strata, which were deposited in outer neritic environments, and Paleocene-Eocene boundary strata of uppermost Zone NP 9 and lowermost Zone NP 10, which were deposited in middle neritic environments, into the shallow subsurface near the western edge of the coastal plain. Upper Paleocene sediments placed in Zone NP 5 through the middle part of Zone NP 9 also occur here ; however, these strata were deposited in shallower, inner to inner middle neritic environments during times of lower sea levels (Gibson, unpubl. data).

Three intermittently cored holes, GL 913, GL 915, and GL 917, and one continuously cored hole at Clayton, all located near the western limit of the present New Jersey coastal plain (Fig. 3), penetrated Paleocene and Eocene strata at drilling depths of 120 m or less. These drillholes penetrated Paleocene-Eocene boundary interval sediments from a deeper water facies than has been observed elsewhere in the Salisbury embayment. It is in these deeper water facies that we find the only known U.S. occurrence of continuous neritic sedimentation across the Zone NP 9-Zone NP 10 boundary.

The thinness and the fine-grained nature of the Paleocene and Eocene sediments, accompanied by their generally high glauconite content, suggest that a relatively low gradient was present during the Paleocene and Eocene in Piedmont areas west of the present outcrop belt. The presence of a low land gradient during times of significantly higher sea levels in parts of the Paleocene and Eocene would have resulted in the shallower water facies of the transgressive-regressive cycles being deposited a considerable distance to the west of the deeper water facies that are now located at the present western edge of the coastal plain. These shallower water facies covered much of the present-day Piedmont but were removed by subsequent erosion (Gibson *et al.*, 1992).

5.3.1. Upper Paleocene Lithostratigraphy

Upper Paleocene deposits that were penetrated in the GL and Clayton drillholes include the Vincentown Formation and the lower part of the Manasquan Formation (Fig. 1). In these drillholes the Vincentown is a greenish to olive-gray, clayey, glauconitic, quartzose, very fine to fine-grained sand (Figs. 10, 11). Glauconite composes 10-60 % of the sand fraction. The formation is massive with a highly bioturbated fabric. The clay mineral suite is dominated by smectite ; a moderate increase in the amount of kaolinite is seen in the uppermost part of the Vincentown (Gibson, Bybell, & Owens, 1993). Shells are present mainly in the lowermost part of the Vincentown and predominately consist of a brachiopod species of *Oleneothyris* and a thick-shelled oyster. The glauconitic quartz sands of the Vincentown Formation, which encompass Zones NP 5 to NP 9 (Bybell, unpubl. data), are similar in lithology, as well as time span, to sediments of the Aquia Formation in Maryland and Virginia. The clayey, glauconitic, quartzose, very fine grained sand that composes the Vincentown in the shallow subsurface of our 4 drillholes is slightly coarser grained than the silt and clay described by Olsson *et al.* (1988) from their shallow subsurface facies of the Vincentown. The lithological transition to the overlying Manasquan Formation begins in the uppermost part of the Vincentown (see 5.3.3).

The Vincentown Formation was considered by Olsson & Wise (1987) and Olsson *et al.* (1988) to represent inner to middle shelf deposition. The foraminiferal assemblages from the Vincentown deposits in the 4 drillholes used in this study also suggest inner to middle neritic depositional environments. Benthonic foraminiferal assemblages in the upper part of the Vincentown have a moderate to high species diversity of 25-45 species and low to moderate percentages of planktonic specimens, ranging from approximately 5-35 %. The abundant benthonic species in most samples from the shallower, inner neritic environments include *Bolivinopsis emmendorferi, Cibicidoides alleni, Cibicides marylandicus, Epistominella minuta,* and *Pyramidina virginiana.* This assemblage includes all the taxa that Olsson & Wise (1987) listed as characteristic of their shallow-shelf assemblages. The deeper, inner middle to middle neritic environments contain increased abundances of *Trifarina wilcoxensis, Tappanina selmensis,* and planktonic specimens. High abundances of *E. minuta, P. virginiana,* and *T. wilcoxensis* in some intervals suggest the presence of a high nutrient water column that probably was accompanied by reduced oxygen levels in the sediments. However, oxygen levels were not so low as to preclude the formation of abundant glauconite.

The *tau* values, which range between 40-400 in most Vincentown samples, also indicate inner neritic depths (Gibson, 1988). Vincentown samples from slightly deeper water environments, as interpreted from the benthonic assemblage, also have higher *tau* values of 500-800 that suggest inner middle to middle neritic environments.

Samples from both the lower and upper parts of the upper Zone NP 9 Vincentown strata contain abundant, diverse, and well-preserved benthonic and planktonic foraminiferal assemblages. Samples from the middle part of upper Zone NP 9 (98.6-105.7 m in the Clayton corehole) (Fig. 10), which contain only a few recrystallized benthonic foraminifers, probably had the remainder of the assemblage removed by dissolution. The calcareous nannofossil diversity also is much reduced in this middle interval ; most assemblages have fewer than 20 species, and some assemblages contain

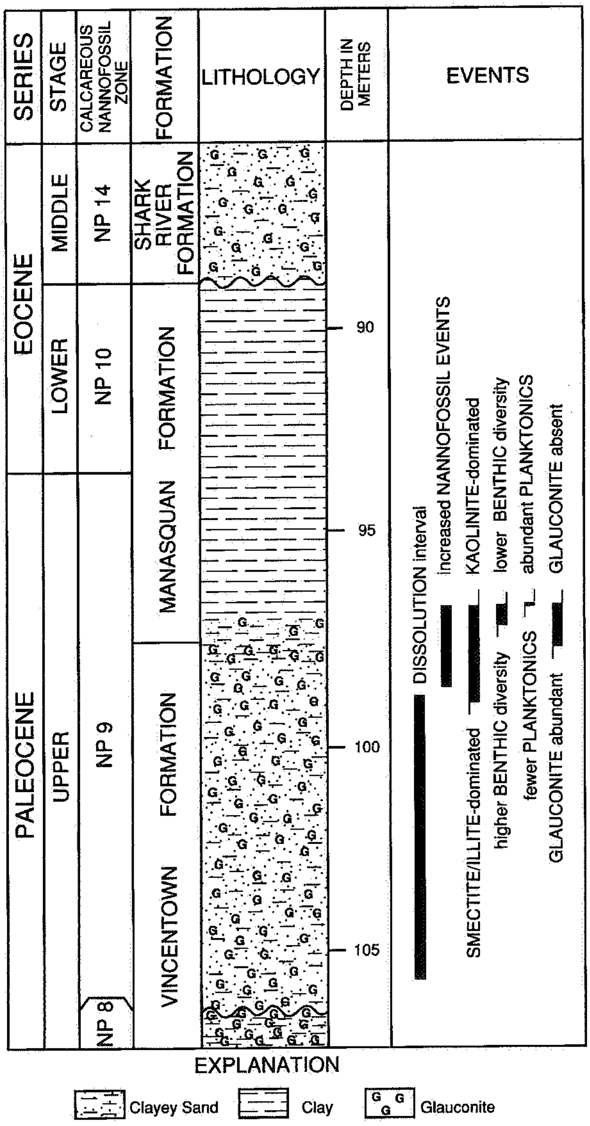

Figure 10. Lithologic section of upper Paleocene to middle Eocene strata in the Clayton, New Jersey corehole.

fewer than 6. This diversity is much lower than the approximately 40 species that occur in upper Zone NP 9 samples taken both above and below this interval. Thinner intervals that are largely or completely barren of foraminifers also are present in a similar position within the upper Zone NP 9 part of the Vincentown in all 3 GL drillholes. Late Paleocene dissolution also is present in a similar position in the Oak Grove corehole in Virginia ; there it occurs in an interval extending downward from about 3 m below the top of the Aquia Formation. The dissolution and recrystal-lization noted in New Jersey and Virginia at this time could result from ground water flow or some other local cause, but it also could be the result of an undetermined global phenomenon.

5.3.2. Upper Paleocene Biostratigraphy

The strata of the Vincentown Formation that are placed in Zone NP 8 contain *Heliolithus riedelii*. The upper strata of the Vincentown that are placed in Zone NP 9 contain *Discoaster multiradiatus* and lack *Tribrachiatus bramlettei*. However, the presence of *Campylosphaera dela* and *Lophodolithus nascens* throughout the entire Zone NP 9 interval (Fig. 12) indicates that these strata belong to the upper fourth of Zone NP 9 (Perch-Nielsen, 1985). Thus, the disconformity between Zone NP 8 and Zone NP 9 in the Vincentown represents a considerable hiatus that involves at least the lower half of Zone NP 9 or approximately 1 m.y.

5.3.3. Paleocene-Eocene boundary unit

A lithologic transition from the clayey, glauconitic sand of the upper part of the Vincentown to the non-glauconitic, slightly silty clay of the overlying Manasquan Formation occurs over a 0.7-m-thick interval in the Clayton corehole (Figs. 10, 11). There is a rapid upward increase in the proportion of dark-olive-gray, bioturbated clay, and a rapid decrease in the proportion of glauconitic silt to very fine grained sand, which commonly occurs as small pods within the clay. Glauconite composes only about 5 % of the silt and sand fraction. There are no shells in this transition interval.

The highly clayey upper part of the transition strata and the overlying dark-olive-gray clay beds are considered to be the lower part of the Manasquan Formation (Fig. 10). The clay contains very little silt and sand (Fig. 11). Mottled, bioturbated, and massively bedded intervals dominate this unit, but there are some intervals that contain thin, dark- and light-colored, sometimes discontinuous, clay laminae. Neither glauconite or mollusk shells are present. This clay unit is 8.2 m thick in the Clayton corehole ; the lower 3.4 m are placed in Zone NP 9 and the upper 4.8 m in Zone NP 10 (Fig. 10). No disconformable surfaces were observed within this homogeneous unit. The clayey sediment continues across the Zone NP 9-NP 10 boundary with no change in grain size (Fig. 11). In the Clayton corehole, the lower Zone NP 10 strata of the Manasquan are disconformably overlain by

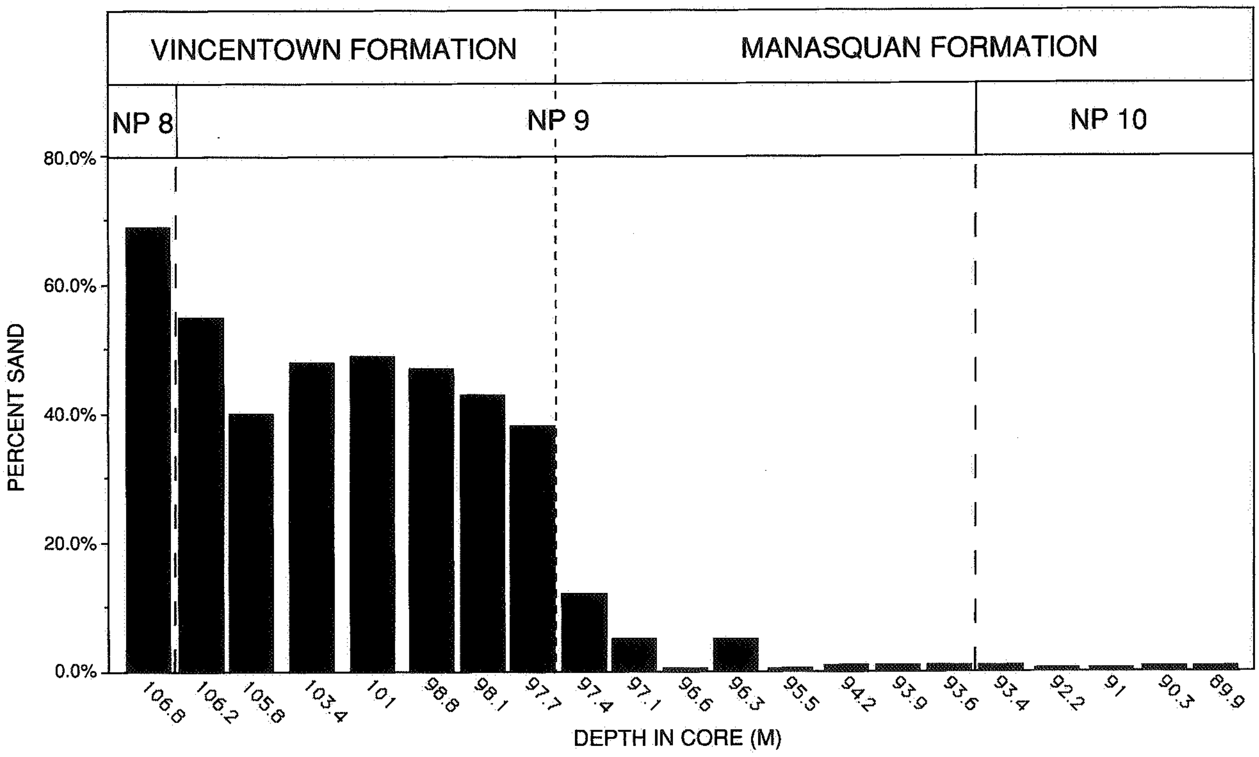

Figure 11. Percent sand in upper Paleocene and lower Eocene strata in the Clayton, New Jersey corehole.

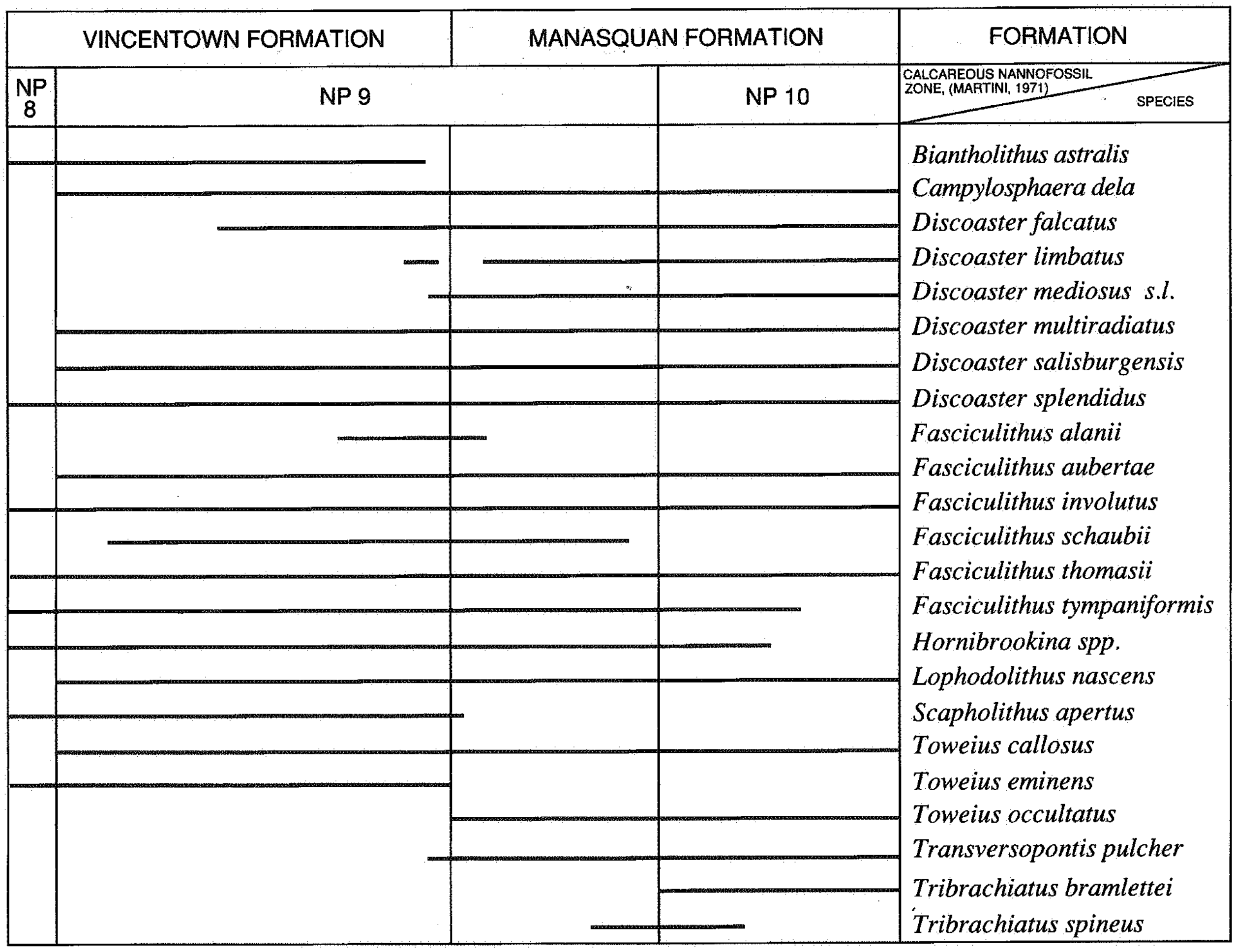

Figure 12. Ranges of selected calcareous nannofossil taxa in upper Paleocene and lower Eocene strata in the Clayton, New Jersey corehole.

glauconitic clayey sand of the lower middle Eocene Shark River Formation (Zone NP 14) (Fig. 10), and in the 3 GL drillholes the Manasquan is overlain by the Kirkwood Formation of Miocene age.

As discussed in Gibson, Bybell, & Owens (1993), associated with the upward lithologic change from the clayey, glauconitic quartz sand of the Vincentown to the clay in the Manasquan is a change in clay mineralogy from a smectite- and illite-dominated suite in the Vincentown to a kaolinite-dominated suite in the Manasquan. The kaolinite domination continues into the lower Zone NP 10 beds. A similar change is seen in the southern and central part of the Salisbury embayment where the Aquia Formation, which is dominated by illite and smectite, is overlain by the Marlboro Clay, which is dominated by kaolinite (Reinhardt *et al.,* 1980 ; McCartan, 1989). Abundant kaolinite continues into the lower part of the overlying Nanjemoy Formation in the Oak Grove corehole (Reinhardt *et al.,* 1980).
Robert & Chamley (1991) and Robert & Kennett (1992) proposed that increased kaolinite in uppermost Paleocene and lowermost Eocene strata in high-latitude areas in the North and South Atlantic was due to increased leaching at that time because of globally warm temperatures and increased precipitation at higher latitudes. The clay mineral data from the entire Salisbury embayment suggest that intensified weathering also occurred at western North Atlantic mid latitudes at this time.

In the Clayton corehole, foraminiferal assemblages in the transitional, glauconitic, sandy clay bed at the base of the Manasquan are similar in number of benthonic species and proportion of planktonic specimens to those occurring in the underlying glauconitic sand of the Vincentown. However, assemblages in the transition bed differ from the underlying Vincentown assemblages by the absence of *Bolivinopsis emmendorferi, Cibicides marylandicus*, and *Cibicidoides alleni* and by the presence of common specimens of *Anomalinoides acutus, Globobulimina* sp., *Pararotalia inconspicua*, and *Pulsiphonina prima*. The differences in the foraminiferal assemblages suggest that oxygen levels were lower during deposition of the basal part of the Manasquan.

In the overlying clay of the Manasquan Formation, glauconite disappears, benthonic species diversity decreases to 15-20 species, and planktonic specimens increase to 53-74 %. The dominant benthonic species include *Anomalinoides acutus, Pararotalia inconspicua, Pulsiphonina prima, Tappanina selmensis*, and *Trifarina wilcoxensis*, along with moderate abundances of *Bolivina* spp., *Bulimina* spp., and *Uvigerina* spp. The composition of the benthonic assemblage, the planktonic percentage, and the absence of glauconite and mollusk shells suggest that these Manasquan beds were deposited in middle to outer neritic environments with low-oxygen bottom conditions. Few changes occur in the common benthonic species across the Zone NP 9-Zone NP 10 boundary, and this indicates that low-oxygen conditions continued at least into the lower part of Zone NP 10. The upper stratigraphic limit of low-oxygen conditions is unknown because in the Clayton and GL drillholes erosional truncation removed the post lower Zone NP 10 strata.

Tau values, which range from 800-1,600 for Manasquan samples, also suggest middle to outer neritic environments. The *tau* values decrease toward the uppermost part of Zone NP 9 and into the basal part of Zone NP 10, and then they increase upwards in the lower part of Zone NP 10 in the Clayton corehole (Fig. 13). This pattern, which also occurs in the GL drillholes, suggests a lowering of relative sea level in uppermost Zone NP 9 that persisted briefly into the lowermost part of Zone NP 10, when relative sea level began to rise. Lowered sea level in the uppermost part of Zone NP 9, if global in extent, helps to explain the absence of similarly aged deposits in many areas of the world.

The decrease in the benthonic species diversity by approximately 50 % and the change in the foraminiferal assemblages to low-oxygen species, which occurs in middle to outer neritic environments in upper Zone NP 9 in New Jersey, are similar to changes reported by Kennett & Stott (1991) in deep-sea Antarctic deposits at approximately the same time. Both studies indicate a change in oceanic and climatic conditions near the end of the Paleocene.

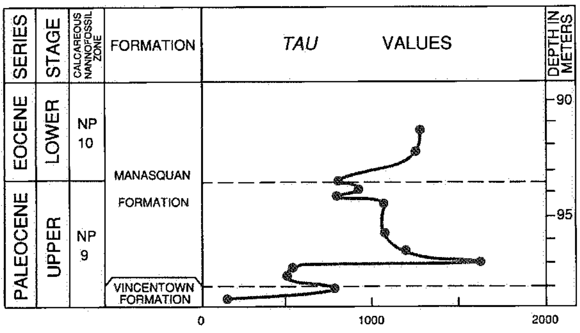

Figure 13. *Tau* values (planktonic % X number of benthonic species) of upper Paleocene and lower Eocene foraminiferal assemblages in the Clayton, New Jersey corehole.

Near the Paleocene-Eocene boundary, deepocean sections show a large-scale extinction of benthonic foraminiferal species (Tjalsma & Lohmann, 1983 ; Miller *et al.*, 1987 ; Thomas, 1989, 1990 ; Kennett & Stott, 1991) in addition to the appearance of lowoxygen species. Although there is a change in the neritic benthonic foraminiferal assemblages to low-oxygen forms within the upper part of Zone NP 9 in New Jersey, few taxa become extinct at that time, and there also is little change in the composition of the lowoxygen assemblages across the Zone NP 9-NP 10 boundary. There are few extinctions of inner neritic benthonic taxa in the upper part of Zone NP 9 and the lower part of Zone NP 10 in the Oak Grove corehole in Virginia (Gibson, Andrews *et al.*, 1980, fig. 6). The most significant benthonic extinction occurred in the Oak Grove corehole between assemblages from Zone NP 8 and NP 9. The most noticeable foraminiferal change near the Zone NP 9-NP 10 boundary in these inner neritic deposits is the first appearance of several benthonic species at or near the base of Zone NP 10 in the Nanjemoy Formation (Gibson, Andrews *et al.*, 1980).

Inner neritic deposits from uppermost Zone NP 9 and lowermost Zone NP 10 are absent due to erosional truncation in SW New Jersey. However, benthonic assemblages from the southern part of the Salisbury embayment in Virginia and Maryland suggest that low-oxygen bottom conditions may have been present even in inner neritic environments during this period of time. The uppermost Zone NP 9 strata of the Manasquan Formation in New Jersey, which contain low-oxygen benthonic assemblages, appear to be coeval with the Marlboro Clay in Maryland and Virginia (Fig. 1). Calcareous foraminiferal assemblages are scarce in the Marlboro, and this makes it difficult to determine with certainty that low-oxygen conditions were present during deposition of this shallow-water unit. However, the calcareous benthonic assemblages presently known from the Marlboro are unusual in their dominance by very high percentages of *Pulsiphonina prima* and *Anomalinoides acutus*. These assemblages suggest that unusual environmental conditions were present even in shallow waters during the latest Paleocene. Several species that first appear in these shallow-water environments in the lower part of the Nanjemoy in lower Zone NP 10 presumably had low-oxygen environmental tolerances or preferences ; they include *Bolivina* sp. C, *Buliminella elegantissima, Florilus* sp. A, *Gyroidinoides octacameratus, Loxostomum* sp. A, and *Turrilina robertsi* ; Gibson, Andrews *et al.*, 1980 ; Gibson, unpubl. data). These assemblages probably are coeval with the deeper-water, low-oxygen assemblages found to the north in lower Zone NP 10 assemblages of the Manasquan.

Olsson & Wise (1987) reported an uppermost Paleocene sequence of gray clay, silt, and fine sand between the Vincentown and Manasquan Formations in New Jersey subsurface sections ; they suggested that deposition of this sequence took place in middle to outer shelf environments. A low abundance, low-diversity benthonic foraminiferal assemblage, usually dominated by *Spiroplectammina spectabilis*, is common in this unit. In one of their drillholes that is located slightly south of our 4 drillholes, *Pulsiphonina prima* and *Tappanina selmensis* are the most commonly occurring species in this unit. Based on planktonic foraminifers and calcareous nannofossils, Olsson & Wise (1987) suggested placement of their unit in the uppermost Paleocene. They considered the sharp changes seen in the benthonic foraminiferal assemblages in this unit to be a result of possible unconformities at the base and top of the unit. Their unit appears to be similar lithologically, in a similar stratigraphic position, and from similar environments to the beds herein placed in the lower part of the Manasquan (upper Zone NP 9 and lower Zone NP 10). They, however, did not report any strata of earliest Eocene age in their unit ; lowermost Eocene strata may have been eroded or may not have been recognized.

The FAD of *Tribrachiatus bramlettei* occurs 3 m above the base of the Manasquan. This species marks the base of calcareous nannofossil Zone NP 10 (Martini, 1971). Although several species have their FAD's and LAD's above or below the Zone NP 9-NP 10 boundary in the New Jersey drillholes (Fig. 12), the FAD of *T. bramlettei*, the zonal marker species, is the only nannofossil event that occurs precisely at the zonal boundary. A more significant number of changes in the calcareous nannofossil flora occur within the upper part of Zone NP 9 at or near the Vincentown-Manasquan contact (Fig. 12) where there are the FAD's or LAD's of 8 taxa ; this interval also contains significant changes in the benthonic foraminiferal fauna, the sporomorph assemblage, and the clay mineral suite.

5.3.4. Lower Eocene Lithostratigraphy and Biostratigraphy

Mountain (1987), Olsson & Wise (1987), and Poag & Low (1987) considered lowermost Eocene strata to be absent from all or most of the outer coastal plain, the shelf, and the slope off New Jersey. Strata of this age do occur in SW New Jersey in our 4 coreholes, and their absence seaward could signify subsequent removal by erosion, which commonly occurs in the Atlantic Margin (Mountain, 1987 ; Gibson *et al.*, 1992). A second possibility is that these lowermost Eocene strata do occur downbasin but in a very thin condensed interval that has not yet been identified because of wide sample spacing or a fragmentary drilling record.

The clay lithology of the lower part of the Manasquan Formation from upper Zone NP 9 and lower Zone NP 10 in our drillholes resembles the lithology described by Olsson & Wise (1987) for the lower Eocene Deal Member in downbasin areas of New Jersey. Our occurrences suggest that a lithology similar to the Deal Member extends into strata of latest Paleocene age in the subsurface. The recognition in New Jersey of lower Eocene deposits belonging in lower and upper Zone NP 10, and Zones NP 11, NP 12, and NP 13 (Poore & Bybell, 1988 ; Bybell, unpubl. data) suggests that portions of the unconformities proposed by Olsson & Wise (1987) for New Jersey are small or nonexistent in the shallow subsurface.

6. CONCLUSIONS

1. A disconformity, representing a short to moderate time interval, is present between uppermost Zone NP 9 (Paleocene) and lowermost Zone NP 10 (Eocene) strata in the eastern Gulf of Mexico Coastal Plain and in the southern to central Atlantic Coastal Plain in the eastern U.S. The hiatus is of greater duration in the eastern Gulf Coastal Plain outcrop belt than in the north-central Atlantic Coastal Plain. The only known record of continuous neritic deposition across the Zone NP 9-NP 10 boundary in the eastern U.S. occurs in shallow subsurface sections in SW New Jersey in the northern Atlantic Coastal Plain.

2. Upper Paleocene deposits of Zone NP 9 age in the eastern Gulf Coastal Plain consist of 2 thick sequences in upper Zone NP 9 (considered to be third-order sequences with thick regressive deposits in the upper part of each sequence) and 2 thinner sequences in the lower part of Zone NP 9 (considered to be less than third order). There are 2 cycles present in upper Zone NP 9 strata in the southern and central parts of the Salisbury embayment, but their age relationship with the Alabama sequences is uncertain.

3. Lower Eocene deposits in the eastern Gulf Coastal Plain contain 6 depositional cycles within Zones NP 10 and NP 11. Four of these cycles are present in the lower half of Zone NP 10 ; a periodicity of 100,000 years is probable for these cycles. The 4 lower cycles have a glauconitic quartz sand at the base that is overlain by carbonaceous clay and silt beds. The upper half of Zone NP 10 and Zone NP 11 contain only 2 thin cycles ; these 2 may represent only a relatively small part of this time interval. In the Atlantic Coastal Plain, only 2 cycles are recognized in Zone NP 10 strata, but 4 cycles are seen in Zone NP 11 strata. The age relationship between the 6 Atlantic and 6 Gulf Coast cycles in Zones NP 10 and NP 11 is uncertain.

4. Several concurrent lithologic and biologic changes are present within the upper part of Zone NP 9 in the continuous section across the Zone NP 9-NP 10 boundary in New Jersey and to a lesser extent in the remainder of the Salisbury embayment. The oceano-

graphic and climatic events recorded in the Salisbury embayment have not been recognized in sediments of the southern Atlantic Coastal Plain or in the eastern Gulf Coastal Plain. However, in these more southerly regions, the uppermost Zone NP 9 and lowermost Zone NP 10 interval is either poorly understood or is missing entirely, and it is impossible at this time to determine how far south along the eastern margin of the U.S. these events extended.

4a. In the continuous New Jersey section, the species diversity of the middle neritic benthonic foraminiferal assemblages decreases by approximately 50 % within uppermost Zone NP 9, and the assemblages change from those indicative of more oxygenated conditions to those indicative of less oxygenated conditions. The low-oxygen assemblages continue with little change into the lower part of Zone NP 10. The few calcareous foraminiferal assemblages found in the shallow-marine deposits of the Marlboro Clay of Maryland and Virginia (upper Zone NP 9) suggest that lowered oxygen conditions may have been present even in inner neritic environments ; low-oxygen conditions are present in inner neritic environments in the lower part of Zone NP 10 (Nanjemoy Formation) in Maryland and Virginia.

4b. In contrast to the significant extinction recorded in the deep-sea fauna at this time, there are few extinctions in the neritic benthonic faunas. The major change in the inner neritic assemblages across the Zone NP 9-NP 10 boundary is the appearance of several new benthonic species in the lower part of Zone NP 10.

4c. Associated with the latest Paleocene appearance of low-oxygen benthonic assemblages in New Jersey was a large increase in the percentage of planktonic specimens in the foraminiferal assemblages. The density of the benthonic component, as measured by the number of specimens per gram of sediment, remained similar ; thus, there was an increase in the planktonic component rather than a decrease in the benthonic one.

4d. In the Salisbury embayment, a lithologic and mineralogic change within the upper part of Zone NP 9 occurs both in middle neritic deposits in New Jersey and in inner neritic to marginal-marine deposits in Maryland and Virginia. The lithology changes from highly glauconitic quartz sand below to largely glauconite-free clay above and indicates the presence of a mud-dominated coast at this time. These clay deposits are gray in color in the middle neritic environments, but are largely pink to red in inner neritic environments ; this color difference suggests that oceanic conditions were more strongly reducing in the deeper neritic environments than in the shallower environments. The clay mineral suite changes in upper Zone NP 9 from smectite and illite domination to kaolinite domination, and this change suggests there was increased weathering due to high temperatures and high precipitation. Robert & Chamley (1991) and Kennett & Stott (1991) also reported increased weathering products at this time in the deep sea record.

5. The only calcareous nannofossil event that occurs precisely at the Zone NP 9-NP 10 boundary is the FAD of the zone-defining species *Tribrachiatus bramlettei*. A few other taxa have FAD's or LAD's some distance above or below this boundary. The extinction of *Fasciculithus* occurs above the base of Zone NP 10, and the extinction of this genus only approximately marks the Zone NP 9-NP 10 boundary. A considerably greater change in the calcareous nannofossils (the FAD's and LAD's of 8 taxa) occurs in the upper part of Zone NP 9 along with other biologic and sedimentologic changes.

6. Erosion, which at least in one area can be linked with tectonic uplift, has removed the more upbasin shallow-water facies of upper Paleocene and lower Eocene sediments in some regions. Uplift of the Piedmont area in New Jersey elevated deeper water deposits that contain continuous deposition across the Zone NP 9-NP 10 boundary into the shallow subsurface. Further to the south in the Salisbury Embayment, there was less uplift and only shallow-water deposits with a disconformity at or near the boundary are present in the outcrop and shallow subsurface.

7. *Tau* values, a water-depth indicator based on foraminiferal assemblages, indicate that water depths increased in upper Zone NP 9, then decreased in the uppermost part of Zone NP 9 and the lowest part of Zone NP 10 before increasing again in Zone NP 10. This relative sea-level lowering across the Zone NP 9-NP 10 boundary may partially explain the widespread absence around the world of uppermost Zone NP 9 deposits. The deepest water, late Paleocene-early Eocene foraminiferal assemblages in the Atlantic Coastal Plain occur in Zone NP 11. This occurrence may be coeval with the first widespread occurrence of calcareous planktonic microfossils in the North Sea area ; if coeval, this would suggest that high sea levels occurred in the North Atlantic area at this time.

8. A dissolution interval is present in the upper part of Zone NP 9 in the New Jersey drillholes discussed in this paper and also in the Oak Grove corehole in Virginia. Planktonic and benthonic foraminifers are absent or badly corroded in this interval, and calcareous nannofossil assemblages have a much reduced diversity.

9. A significant change in the sporomorph assemblage occurs in the upper part of Zone NP 9 within the Marlboro Clay. Thirteen taxa become extinct at this time, and numerous taxa make their first appearance in the central Atlantic Costal Plain ; climatic change is

the probable cause.

10. There was a significant climatic change in the latest Paleocene within the upper part of Zone NP 9 that caused warmer oceanic and terrestrial temperatures, increased precipitation, and lower oxygen oceanic conditions. These changes affected the diversity and composition of neritic benthonic foraminifers, caused evolutionary changes in the calcareous nannoplankton and the terrestrial flora, and changed the type of clay mineral produced in mid-latitude, western North Atlantic areas.

11. The Zone NP 9-NP 10 boundary is now placed at 55 Ma, and significant late Paleocene biologic and lithologic changes occurred at approximately 55.2 Ma.

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