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100 Myr record of sequences, sedimentary facies and sea level change from Ocean Drilling Program onshore coreholes, US Mid-Atlantic coastal plain

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ABSTRACT

We analyzed the latest Early Cretaceous to Miocene sections (~110–7 Ma) in 11 New Jersey and Delaware onshore coreholes (Ocean Drilling Program Legs 150X and 174AX). Fifteen to seventeen Late Cretaceous and 39–40 Cenozoic sequence boundaries were identified on the basis of physical and temporal breaks. Within-sequence changes follow predictable patterns with thin transgressive and thick regressive highstand systems tracts. The few lowstands encountered provide critical constraints on the range of sea-level fall. We estimated paleowater depths by integrating lithofacies and biofacies analyses and determined ages using integrated biostratigraphy and strontium isotopic stratigraphy. These datasets were backstripped to provide a sea-level estimate for the past ~100 Myr. Large river systems affected New Jersey during the Cretaceous and latest Oligocene–Miocene. Facies evolved through eight depositional phases controlled by changes in accommodation, long-term sea level, and sediment supply: (1) the Barremian–earliest Cenomanian consisted of anastomosing riverine environments associated with warm climates, high sediment supply, and high accommodation; (2) the Cenomanian–early Turonian was dominated by marine sediments with minor deltaic influence associated with long-term (10^7 year) sea-level rise; (3) the late Turonian through Coniacian was dominated by alluvial and delta plain systems associated with long-term sea-level fall; (4) the Santonian–Campanian consisted of marine deposition under the influence of a wave-dominated delta associated with a long-term sea-level rise and increased sediment supply; (5) Maastrichtian–Eocene deposition consisted primarily of starved siliciclastic, carbonate ramp shelf environments associated with very high long-term sea level and low sediment supply; (6) the late Eocene–Oligocene was a starved siliciclastic shelf associated with moderately high sea-level and low sediment supply; (7) late early–middle Miocene consisted of a prograding shelf under a strong wave-dominated deltaic influence associated with major increase in sediment supply and accommodation due to local sediment loading; and (8) over the past 10 Myr, low accommodation and eroded coastal systems were associated with low long-term sea level and low rates of sediment supply due to bypassing.

INTRODUCTION

The New Jersey coastal plain of the Atlantic passive continental margin provides a 100+million year record of sea-level changes (Miller *et al.*, 2005). Early studies recognized numerous Late Cretaceous to Miocene (~100–7 Ma) transgressive-regressive pulses (Owens & Sohl, 1969; Olsson, 1975). More recent studies placed these transgressions and regressions into a sequence stratigraphic framework (Olsson, 1991; Sugarman *et al.*, 1993). Continuous coring by the Ocean Drilling Program (ODP), both offshore (Legs 150 and 174A; Mountain *et al.*, 1994; Austin *et al.*, 1998) and onshore (Legs 150X and 174AX; Miller *et al.*, 1994, 1996a, b, 1998b, 1999, 2001, 2003b, 2006; Sugarman *et al.*, 2005b, 2004), has provided one of the best dated records of Myr-scale sequences (unconformity bounded units) and a well-defined history of water depth changes for the past 100 Myr. New coreholes allow us to extend this record back to ~110–120 Ma. Backstripping of this well-dated water depth record, by progressively removing the effects of compaction, loading, and thermal subsidence, has provided a sea-level estimate (as discussed in Kominz *et al.*, 1998, 2008; Kominz & Pekar, 2001; Van Sickel *et al.*, 2004; Miller *et al.*, 2005).

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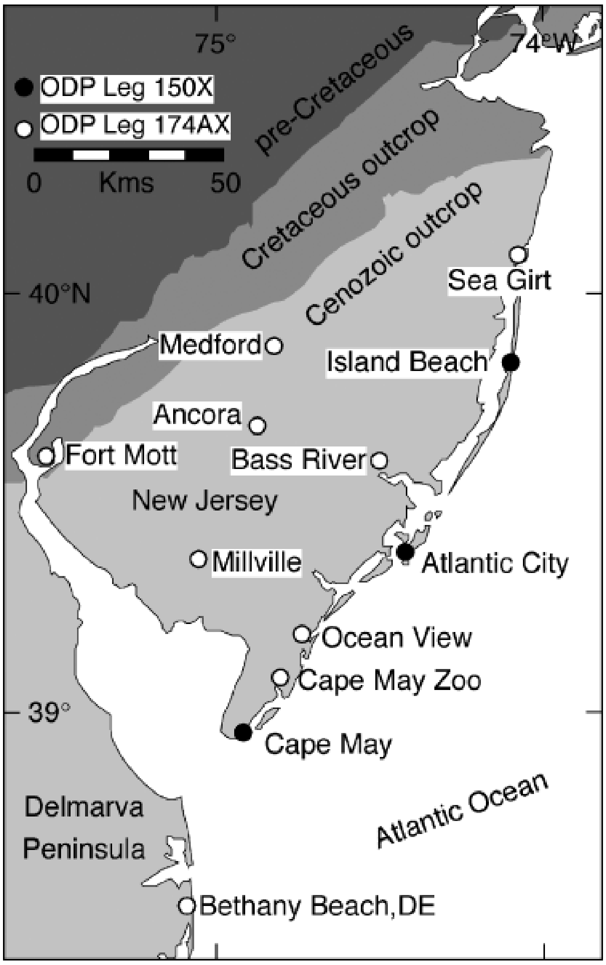

Fig. 1. Location map showing 11 ODP coreholes analyzed as a part of the NJ/Mid Atlantic (MAT) sea-level transect. The recently drilled corehole from Medford, NJ is also shown.

The 11 onshore coreholes drilled in New Jersey and Delaware (Fig. 1, ODP Legs 150X and 174AX) provided the basis for the backstripped sea-level estimates because offshore ODP sites are either in too deep water (e.g. Leg 150 and 174A continental slope sites) or the sediments sampled were restricted to the past 10 Myr (outer shelf Leg 174A sites). Fifteen to seventeen Late Cretaceous and 39–40 Cenozoic Myr-scale sequences were identified on the basis of physical and temporal breaks. The sequences were dated by integrating Sr-isotopic stratigraphy (Late Cretaceous and late Eocene through Miocene), magnetostratigraphy (Paleocene–Oligocene), nannofossil biostratigraphy (Late Cretaceous through Late Eocene), planktonic foraminiferal biostratigraphy (Late Cretaceous through lowermost Miocene), pollen biostratigraphy (Cenomanian through Turonian and middle through upper Miocene) and pollen biostratigraphy (middle through upper Miocene) to provide age resolution better than 1 Myr for most intervals (Miller *et al.*, 1998a, 2004, 2005). Water depth changes within sequences were estimated from lithofacies and biofacies analyses, providing a chronology of water depth changes (Miller *et al.*, 2005).

We used the onshore ODP coreholes to derive a preliminary sea-level estimate by backstripping the corehole records, progressively accounting for the effects of compaction, loading, and simple thermal subsidence (Van Sickel *et al.*, 2004; Miller *et al.*, 2005). In the absence of regional or local tectonics, backstripping provides a eustatic estimate with the greatest uncertainty resulting from errors in water depths assigned. Backstripped records from the onshore ODP coreholes generally yielded similar sea-level estimates (see Kominz *et al.*, this volume, for detailed discussion), suggesting that we have identified regionally consistent relative sea-level changes. The New Jersey margin sequences correlate with those from other passive margins and epicontinental seas (e.g., Browning *et al.*, 1996; Sahagian *et al.*, 1996; Miller *et al.*, 2004) and the oxygen isotope proxy for glacioeustasy (Miller *et al.*, 1998a, 2005), suggesting that global sea-level changes were a dominant process controlling Myr-scale sequences on this margin. Nevertheless no one location can be used to provide a eustatic record, and the global sea-level record of Miller *et al.* (2005) should be regarded as a preliminary, testable model with shortcomings that include limited Late Cretaceous coverage (only two coreholes were used in the synthesis), uncertainties in water depth estimates, and other limitations (Kominz *et al.*, this volume).

Although eustasy appears to be a dominant control on Myr-scale sequences (Miller *et al.*, 1998a, 2005), facies changes within sequences reflect changes in accommodation (including effects of sea level and subsidence) and sediment supply and redistribution (e.g. Posamentier *et al.*, 1988; Reynolds *et al.*, 1991; Swift & Thorne, 1991; Swift *et al.*, 1991). In addition, the general depositional settings (e.g. carbonate ramp vs. siliciclastic; prograding deltaic vs. storm dominated shelf; referred to as the depositional phase) on the New Jersey margin shifted on a 10 Myr-scale in response to changes in accommodation, shifts in sediment supply and provenance, and long-term sea-level changes (e.g. Owens & Sohl, 1969; Poag & Sevon, 1989). Tectonics also plays a role, even on a passive continental margin like the mid-Atlantic margin (Browning *et al.*, 2006). The challenge to stratigraphers has been to disentangle the effects of sea level, subsidence, and sediment supply from the stratigraphic record. By assuming that the sea-level estimates of Miller *et al.* (2005) and Kominz *et al.* (this volume) provide a reasonable first-order approximation to eustasy, we can begin to truly evaluate the role of tectonics and sediment supply on the molding of sequences.

In this contribution, we revisit Upper Cretaceous to Miocene sequences drilled on the US Mid-Atlantic (New Jersey and Delaware) coastal plains (Fig. 1) and provide a history of sequences from $\sim$110 to 7 Ma (Fig. 2). We discuss changes in facies successions (Fig. 3), illustrate critical lowstand deposits (Fig. 4), delineate facies models for terrestrial (Figs 5 and 6), deltaic (Fig. 7), and shoreface environments (Fig. 8), and illustrate the evolution of depositional phases against long-term sea-level estimates (Figs 9–12). This comparison allows us to evaluate the impact of long-term sea-level changes on depositional phases. The

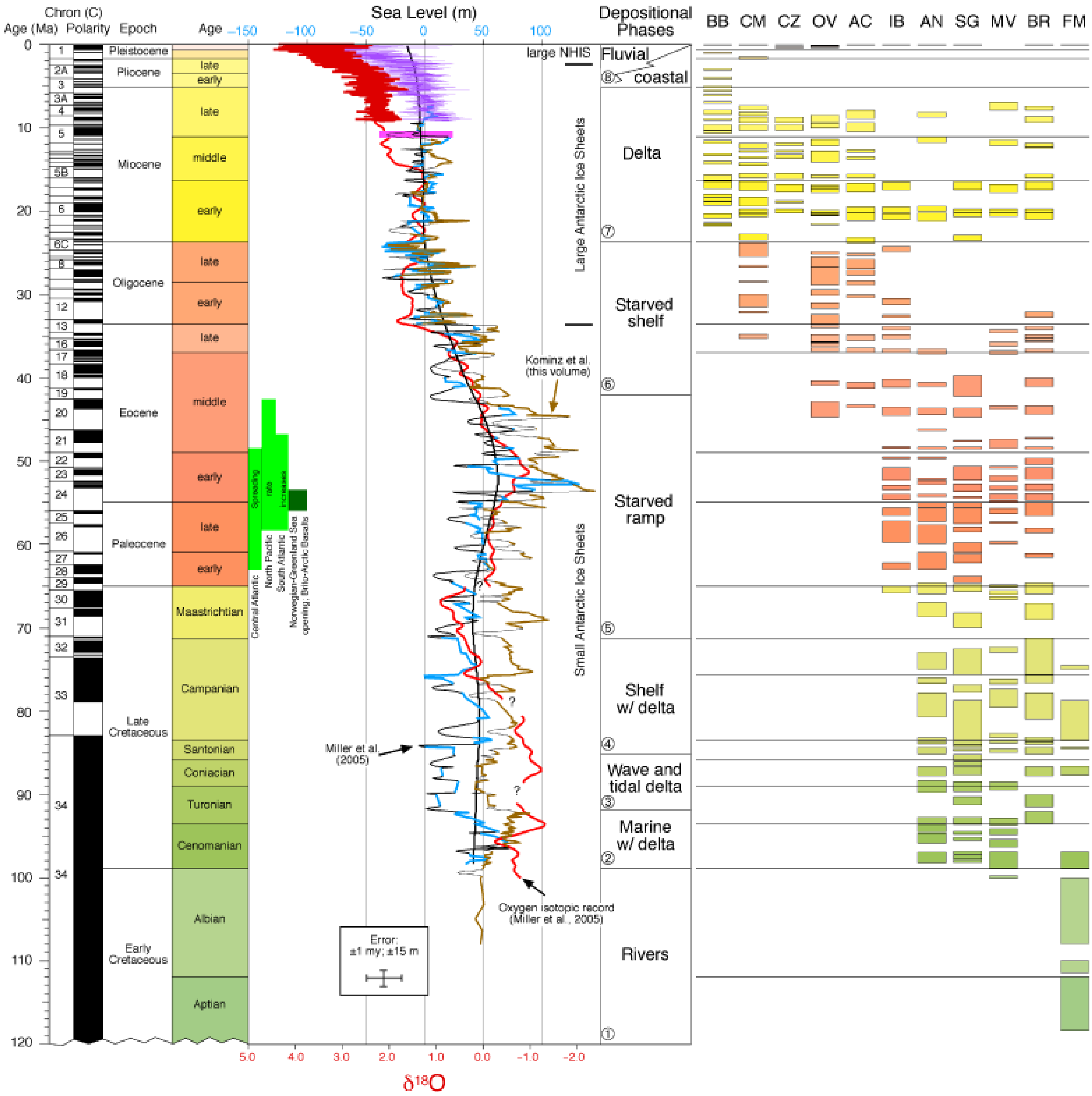

Fig. 2. Distribution of sediments in sequences as a function of time. Sea level curve in blue from Miller *et al.* (2005). Sea level curve in brown from Kominz *et al.*, (this volume). Red oxygen isotopic curve from Miller *et al.* (2005). Depositional phases are described in the text BB, Bethany Beach core; CM, Cape May core; CZ, Cape May Zoo core; OV, Ocean View core; AC, Atlantic city core; IB, Island Beach core; AN, Ancora core; SG, Sea Girt core; MV, Millville core; BR, Bass River core; FM, Fort Mott core; NHIS, Northern Hemisphere Ice Sheets.

sediments provide information on accumulation rates, provenance and margin current phase. By integrating these data, we can demonstrate the relative effects of the triad of processes on the evolution of this margin: sea level, subsidence, and sediment supply.

IDENTIFICATION OF SEQUENCES AND SYSTEMS TRACTS

Sequence boundaries are recognized in cores on the basis of physical stratigraphy. Significant features include: irregular contacts, reworking, rip-up clasts, bioturbation, major facies changes, gamma ray peaks (often indicating lag deposits), and age breaks. Lithofacies were recognized using textural and structural data (grain size, general lithology, bedding, and sedimentary structures), quantitative measurement of the medium–coarse sand, very fine–fine sand, and silt/clay fractions, and semi-quantitative counts of glauconite, carbonate grains, and mica in the sand fraction (Figs 9–12).

Facies changes within sequences generally follow a repetitive transgressive-regressive pattern (Fig. 3). The major elements in this pattern are, from the base upward:

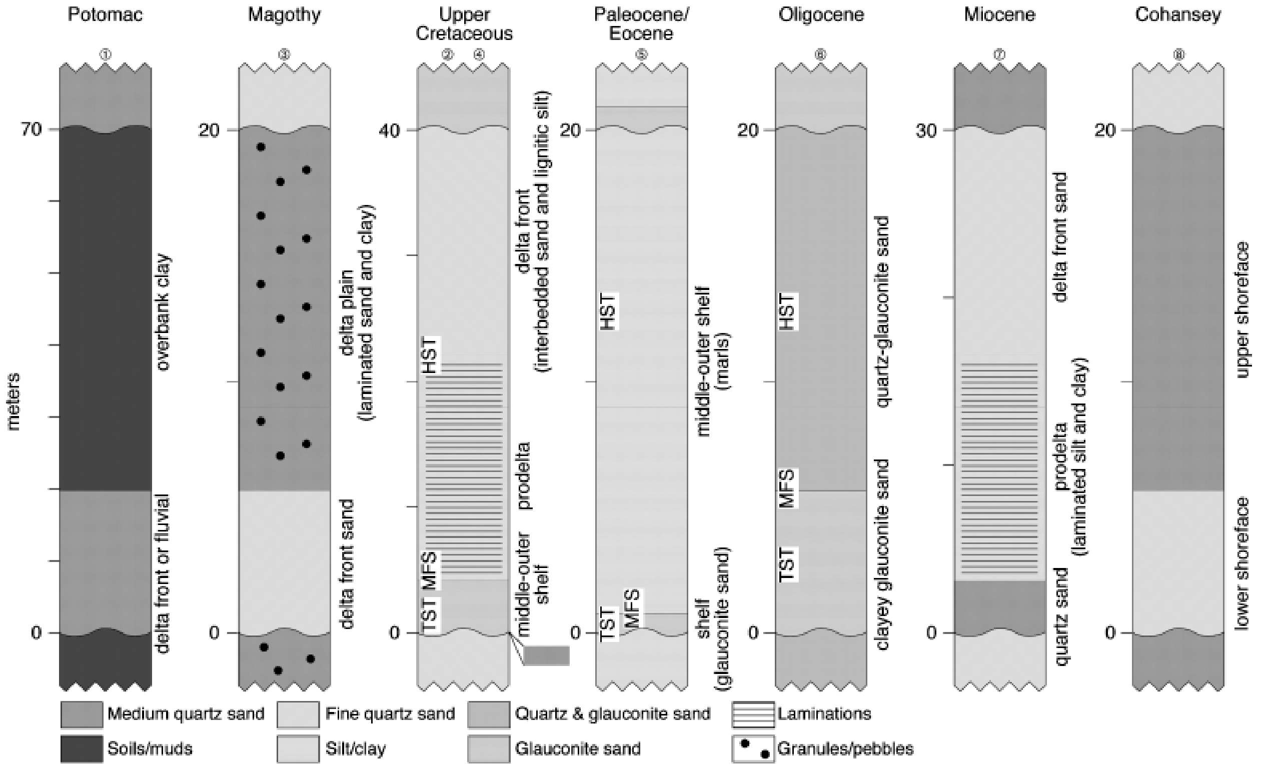

Fig. 3. Diagrams showing facies patterns in typical sequences from different depositional phases. TST, transgressive systems tract; HST, highstand systems tract; MFS, maximum flooding surface. Sequence thicknesses are based on the following sequences: Potomac–Unit 2 at Fort Mott corehole; Magothy–Unit 3 at Sea Girt corehole; Upper Cretaceous–Marshalltown sequence at Bass River corehole; Paleocene/Eocene–Sequence E7 at Ocean View corehole; Oligocene–Sequence O5 at Ocean View corehole; Miocene–sequence Kwla at Ocean View corehole; Cohansey–Cohansey sequence at Cape May Zoo corehole. Numbers at the tops of lithologic columns refer to major phases of sedimentation identified in the text.

(1) a basal unconformity (lower sequence boundary); (2) a generally thin lower sand (glauconite in the Cretaceous to Paleogene; quartz in the Miocene) that is assigned to the transgressive systems tract (TST; Posamentier *et al.*, 1988); marginal marine and terrestrial sequences [e.g., Magothy Formation] typically lack the glauconite sands; (3) maximum water depth, indicative of the maximum flooding surface (MFS), is generally found within the upper glauconitic sands and is identified by benthic foraminfers, a change from transgressive to regressive facies and a gamma log increase; (4) a coarsening-upward succession of regressive silt [lower portion of the highstand systems tracts (HST) of Posamentier *et al.*, 1988] which includes the early fall of relative sea-level; (5) upper quartz sand of the upper HST; and (6) the upper sequence boundary. Facies generally deepen by overstepping across coastal plain sequence boundaries from inner neritic/delta front at the top of the underlying sequence to middle neritic at the base of the overlying sequence. Sequences from the New Jersey coastal plain often yield a distinctive gamma-log signature, with high values at sequence boundaries, in glauconite sands, and at flooding surfaces, moderate values in silty clays, and low values in the quartz sands (Lanci *et al.*, 2002). Eocene sequences (Fig. 3) generally lack HST sands due to their greater depths of deposition (middle–outer neritic) and reduced sediment supply. Oligocene sequences (Fig. 3) have a mixture of quartz and reworked glauconite sand in the HST (Pekar *et al.*, 2000).

Because transgressive (deepening upward in cores) deposits interpreted as TSTs are thin and lowstand systems tracts largely absent (i.e., glauconites occur immediately above sequence boundary unconformities), transgressive ravinement surfaces (TS) are concatenated with sequence boundaries. Flooding surfaces, particularly MFSs, may be differentiated from sequence boundaries by the association of erosion, rip-up clasts and age breaks discerned from Sr-isotopic and bio-stratigraphy at sequence boundaries, and lithofacies successions and benthic foraminiferal biofacies indicating a shift from transgression to regression at MFS. Both SBs and MFS are often associated with hot zones on gamma logs; the former because of hiatuses associated with lag deposits (phosphorites, uranium-rich zones) and the latter because of slow deposition and accumulation of clays.

A few lowstand systems tracts (LST) were encountered that provide critical constraints on the range of sea-level fall. These LSTs include the basal part of the Navesink sequence (Cretaceous) at outcrop and in the Sea Girt corehole (Fig. 3a in Miller *et al.*, 2004), the Marshalltown sequence (Cretaceous) in the Sea Girt corehole (Fig. 4) and a Kirkwood (Kw2b) sequence in the Ocean View and Cape May Zoo coreholes (Miocene). Where encountered,

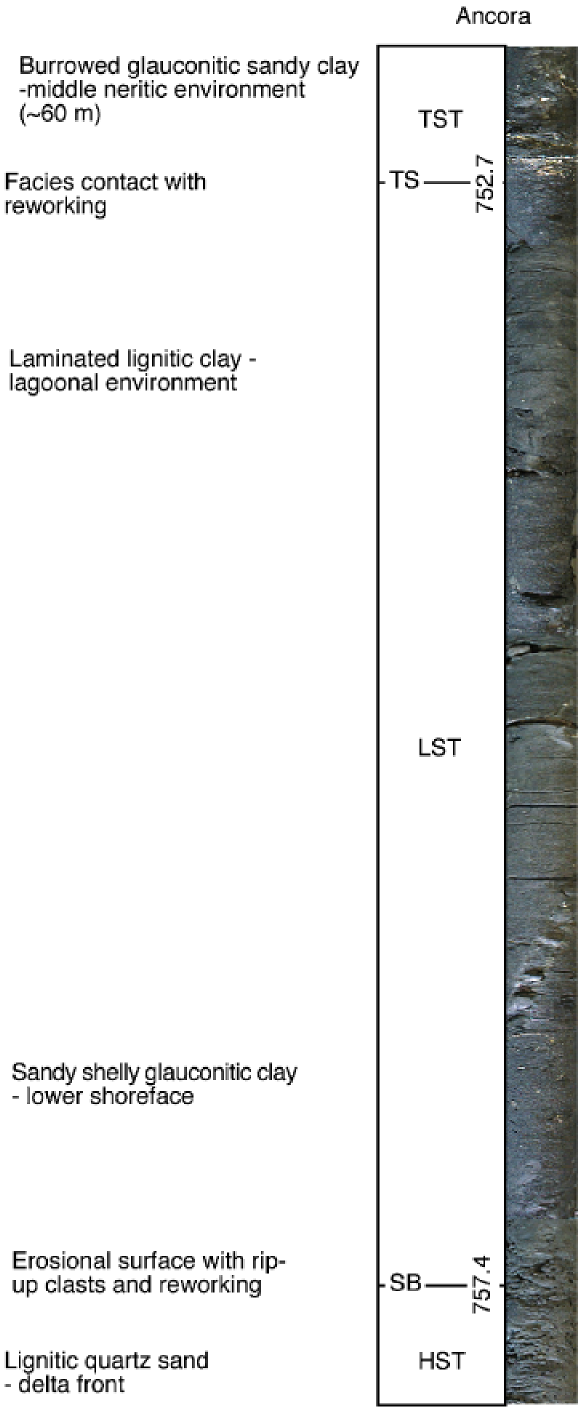

Fig. 4. Lowstand systems tract from the Marshalltown sequence in the Sea Girt and Ancora coreholes. TST, transgressive systems tract; LST, lowstand systems tract; HST, highstand systems tract; TS, transgressive surface; SB, sequence boundary. Numbers refer to depth in the core in feet and meters.

LSTs are thin (<2 m) and shallow upsection to the TS above which lithofacies indicate deepening upward (Fig. 4). This change in stacking pattern above an erosional sequence boundary is interpreted as a LST. The LSTs, probably representing incised valley fill, often contain rip-up clasts or reworked pods of the underlying lithology [e.g. 691.2–687.1 ft (209.4–210.7 m) at Sea Girt, Fig. 4]. For example, in outcrop, the LST of the Navesink sequence consists primarily of sand deposited within a silty matrix. The few observed LSTs are important because they allow us to sample a nearly full range of the total sea-level cycle.

Continuous coring at both along strike and up and down dip locations has allowed us to sample and identify a greater number of sequences than could be identified in outcrop or in single coreholes on the coastal plain. For example, outcrop studies typically recognize five to seven Late Cretaceous transgressive–regressive cycles (Olsson, 1963, 1975; Owens & Sohl, 1969) that are now recognized as sequences (Olsson, 1991). Our initial studies of the Ancora and Bass River coreholes (Fig. 1; Miller *et al.*, 1998b, 1999, 2004) resulted in identification of 11–14 Late Cretaceous sequences. Integration of that record with new corehole records from Millville (Sugarman *et al.*, 2005b) and Sea Girt (Fig. 1; Sugarman *et al.*, 2004), and well log correlation from regional drillholes (e.g. Kulpecz *et al.*, 2008), has expanded the number of Upper Cretaceous sequences to the 15 shown here (Figs 2, 9 and 10) plus the possibility of two additional Coniacian Magothy sequences (Kulpecz *et al.*, 2008). Virtually all of the sequence boundaries have recognizable hiatuses and can be traced regionally, indicative of lowering of base level associated with a sequence boundary. We recognize 7 Myr-scale Paleocene sequences, 12 Eocene sequences, eight Oligocene (seven from Pekar & Miller, 1996; Pekar *et al.*, 2002 recognized one additional earliest Oligocene sequence, ML from the ACGS#4 corehole drilled before ODP 150X; Pekar *et al.*, 2000), and 14 Miocene sequences (not including possible subdivisions of the Kwla and Kw2 into higher order sequences). Cretaceous sequences are named after their basal lithologic units (Miller *et al.*, 2004), typically a glauconite sand (e.g. the Marshalltown (Ma) sequence; Figs 9 and 10). Paleocene sequences (Fig. 10) are named Pa0 through Pa3b oldest to youngest (nomenclature modified after Liu *et al.*, 1997 to reflect the identification of three additional sequences), Eocene sequences (Figs 10 and 11) are named E1–E11 (Browning *et al.*, 1996), Oligocene sequences (Fig. 11) are named O1–O6, and Miocene sequences (Fig. 12) are named after the Kirkwood and Cohansey Formations (Kw0–Kw3; KwCh1–KwCh6).

FACIES MODELS

The sediments of the New Jersey coastal plain were deposited in a wide range of sedimentary environments. Deltaic and delta-influenced deposition has dominated much of the last 120 Myr of New Jersey sedimentation, but we also encountered sediments from fluvial environments, wave/storm-dominated shorelines, and neritic environments. We used facies models (Figs 5, 7 and 8) to identify the sedimentary environments in which the sediments accumulated.

We recognized three broad environments of deposition within our coastal plain coreholes and developed facies models to help identify their subenvironments (Figs 5, 7

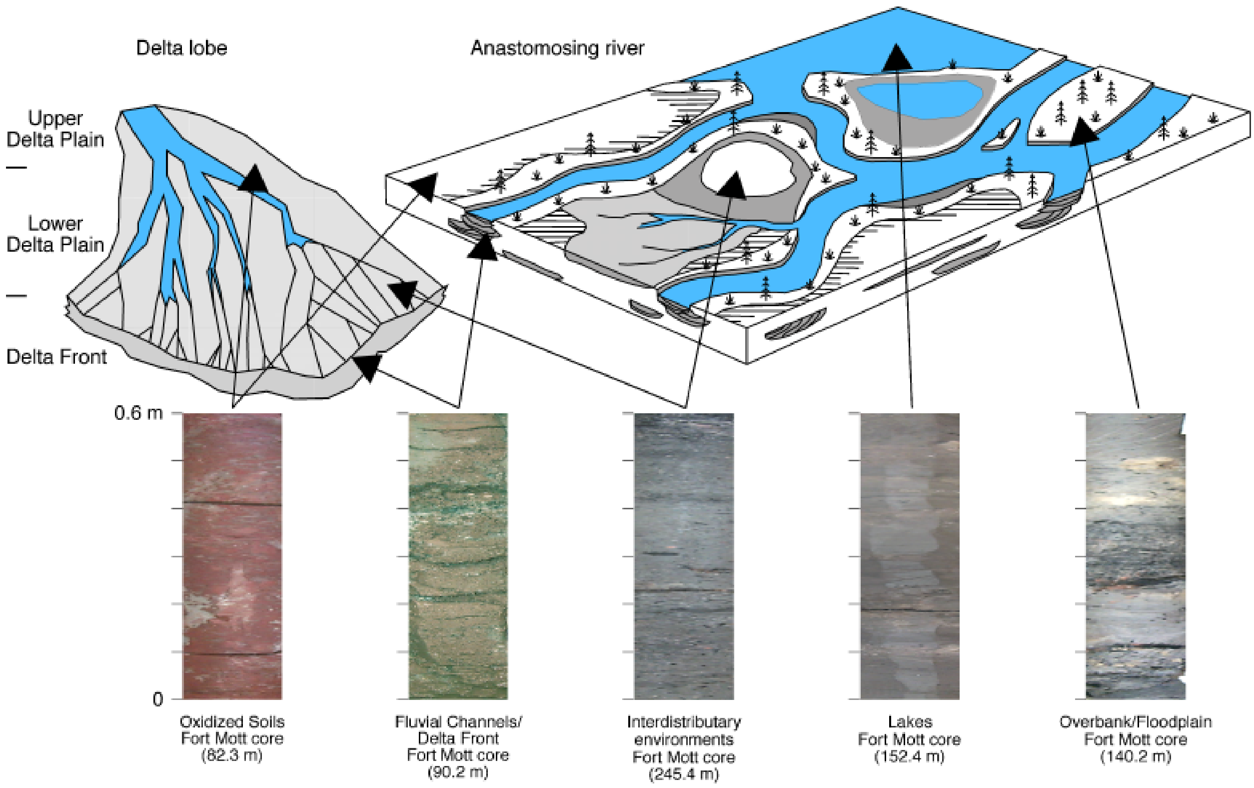

Fig. 5. Facies model for anastomosed river sedimentation (figure modified from Miall, 1996) and delta plain sedimentation (figure modified from Elliott, 1986). Photographs are examples of each facies in core from given levels in the Fort Mott borehole.

and 8). Terrestrial environments are those above mean high tide and the sediments accumulated without marine influence. Deltaic environments include lower delta plain, delta front and prodelta environments in which sediments accumulated as a result of a delta building into marine environments. Wave-dominated-shoreface and shelf sediments accumulated in a marine environment without evidence for a deltaic source. Below, we describe criteria for each environment and then where these criteria were observed in our cores to make interpretations of paleoenvironments.

Terrestrial criteria

Terrestrial paleoenvironments in our New Jersey coastal plain coreholes are characterized chiefly by fluvial channel sands and well-developed soil profiles deposited on floodplains that accumulated on a heavily vegetated delta plain (Owens & Sohl, 1969; Sugarman *et al.*, 2004). Most terrestrial delta plains are extensive lowlands with active and abandoned distributary channels (Allen, 1970). The delta plain is dominated by fluvial processes and is typically divided into two distinct subenvironments: the upper delta plain and the lower delta plain. The upper delta plain lies above high tide and is not influenced by the ocean and thus contains fresh water deposits. Lower delta plains are affected by fluvial and tidal processes and thus contain brackish water deposits and are not terrestrial. Lower delta plain deposits were not confidently identified in part because of the low tidal range in the region.

Two major facies are found in an upper delta plain. Subordinate channel-fill facies consist of gravel and coarse-grained sand (Smith & Smith, 1980). Fine-grained overbank deposits comprise the majority of the sediments. Levees flanking the rivers consist of alternations of sandy silt and silty sand containing roots. Low areas between levees have lakes and peat bogs or back swamps; they are differentiated based upon the frequency with which they receive water and their connection to the main channels. Lakes are intermittently supplied with fine sediment including laminated clay and silty clay with sparse organic matter. Peat bogs and back swamps have no direct connection to the main channels and may receive clastic sediments (mainly organic-rich silty clay and clayey silt) only during times of flood. Finally, crevasse splays are recognized as thin layers of sand and fine gravel rarely more than 40 cm thick (Smith & Smith, 1980).

Many modern workers consider the channels in an upper delta plain to be fundamentally similar to anastomosing channels [see Makaske (2001) for a discussion]. Like braided systems, anastomosed river systems have bars separating the channels. Anastomosed systems differ from braided systems in that their channel and bar stability prevents the river from reworking organic-rich sediments. As a result, anastomosed systems are dominated by fine floodplain deposits and organic-rich sediments or coals (Makaske, 2001). We follow the classification of major environments and sedimentary facies for anastomosing river systems described by Smith & Smith (1980). Anastomosed river systems consist of multiple coexistent

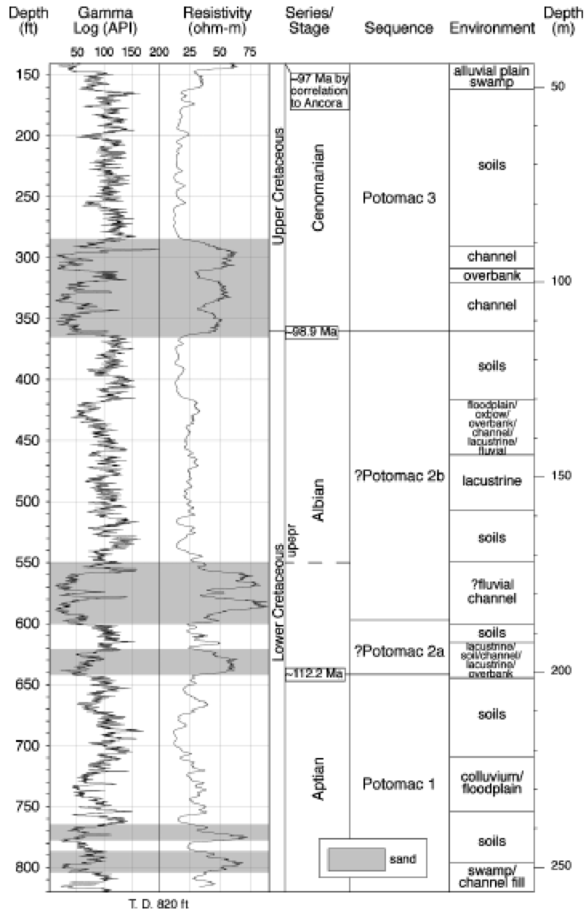

Fig. 6. Sequences and environments for units recovered from the Fort Mott borehole. Gray areas in the gamma and resistivity log columns indicate aquifers.

channels separated by stable islands or bars (Smith & Smith, 1980). Anastomosed systems develop in regions experiencing rapid sea-level rise such as the Holocene transgression (Makaske, 2001).

Terrestrial observations

The Fort Mott corehole is the only continuous corehole through the Potomac Formation (Table 1) studied to date (Fig. 1). It provides examples of the facies and depositional environments of this terrestrial unit that are supported by comparison with outcrop and other limited subsurface studies (a corehole at Medford, NJ, drilled in May 2007 extensively sampled the Potomac Formation). The Potomac Formation in the Fort Mott borehole (Figs 5 and 6) is dominated by mottled red and bluish gray sediments. The sediments often include sphaerosiderite nodules interpreted to have formed by soil processes (Retallack, 1990). These sediments represent paleosols produced in an overbank setting with the different colors indicating varying degrees of oxidation during soil development (Retallack, 1990). Gleyed soils, more typical of marshy wetlands, are bluish gray, due to waterlogging, and the resultant loss of iron compounds and oxygen. Red lateritic soils form under conditions of stability and intense weathering [e.g., 259.4–274.8 ft (79.1–83.8 m) in the Fort Mott corehole; Fig. 6]. The reddish clays represent soils that developed on overbank deposits on the banks of streams and in between adjacent channels. Gray clays were deposited in small lakes and as marsh deposits.

Coarser clastic material sampled consists of several distinct fining upwards sequences including: (1) pebbly granule beds at the bottom grading upwards into fine-medium grained sands containing opaque black mineral laminae, and (2) gravelly, very coarse sand overlain by flat-laminated to cross-laminated, fine–medium grained quartz sand with whitish, silty, very fine sand with silt lamina on top. Fining-upward successions represent various sized fluvial channels. The pebbly granules grading upwards to fine–medium grained sands represent smaller channels. Three such stacked channels were encountered from 300.4 to 310.6 ft (91.6–94.7 m) at Fort Mott. The gravelly, very coarse sand and flat- to cross-laminated, fine quartz sand topped by very fine sand suggest a larger channel [e.g. 290.65–297.3 ft (88.6–90.6 m) Fig. 6].

Several other sedimentary associations occur in the Potomac Formation at Fort Mott. Dark gray, slightly silty clay and clayey silt commonly contain scattered small charcoal fragments and thin micaceous sand interbeds [e.g. 445–452 ft (135.6–137.8 m) and 615.4–620.1 ft (187.6–189.0 m) Fig. 6]. The sediments are often laminated with scattered dark (organic-rich) laminae. The clays lack the sphaerosiderite nodules found in the mottled red and bluish gray sediment described above. Also encountered were interlaminated silty clay and very fine sand [e.g. 154.3–164.2 ft (47.0–50.0 m) Fig. 6] containing scattered cross-laminations, deformed laminae, dark gray organic-rich beds and scattered charcoal. In addition, some facies are dominated by very charcoal-rich, micaceous, very fine sand, sandy silt and sandy charcoal beds with cross beds and soft sediment deformation features. These facies represent lacustrine, swamp, and crevass splay subenvironments. The dark gray, silty clay and clayey silt with charcoal and micaceous sand interbeds suggests a lacustrine environment. The lack of sphaerosiderite nodules typical of Upper Cretaceous soils deposited in a megathermal climate (e.g. White *et al.*, 2001) argues against this facies being another soil deposit. Laminated, fine-grained (clayey silt to silty clay) sediments lacking sand, abundant organic matter or pyrite argues against oxbow lake subenvironments, and indicates deposition in larger, floodplain standing lakes [e.g. 461–508.15 ft (140.5–154.9 m) Fig. 6]. A swamp environment is interpreted for the interlaminated clay and sand containing dark gray organic-rich beds and scattered charcoal. Finally crevasse splays are recognized from the charcoal-rich, sandy material. The cross beds within these sediments indicate deposition in a moving current but the presence of charcoal, mica and silt point towards limited winnowing of the sediment after deposition.

We note the predominance of fine, overbank and lacustrine deposits that are more typical of anastomosed versus braided or meandering channel systems (Makaske, 2001).

Lakes are a common subenvironment of such systems (Makaske, 2001), and we interpret numerous organic-rich, fine-grained sections as fluvial-lacustrine deposits (Sugarman *et al.*, 2004). These lacustrine sediments include not only oxbow lake deposits, but also more regionally pervasive lake deposits with a distinct cyclicity (Sugarman *et al.*, 2004). Although lakes are important, the dominant subenvironments of modern anastomosed systems are overbank/levee soils and subaqueous swamps (Smith & Smith, 1980), consistent with our interpretation of the Potomac Formation at Fort Mott (Sugarman *et al.*, 2004). Further coring is needed to determine the extent of this system

Delta front-prodelta and estuarine criteria

The delta front is the zone just seaward of the river mouth where river currents enter the basin and the sediments are further dispersed by basinal processes (Allen, 1970; Bhattacharya & Giosan, 2003). The energy regime of the delta front is used to define the delta type. River-dominated deltas (e.g. Mississippi delta) are characterized by little sediment reworking from waves or tides (Allen, 1970; Bhattacharya & Giosan, 2003). These deltas build lobes into the sea and little more than the levees of the distributary channels are exposed above sea level. The interdistributary bays are filled with mud and crevasse splay deposits. Progradation of the distributary channel and its mouth bar into the open sea produces a coarsening upward succession. Abandonment of the channel causes the sequence to be overlain, often abruptly, by mud (Allen, 1970; Bhattacharya & Giosan, 2003).

Wave-dominated deltas form where significant wave action reworks the delta front sediments (Allen, 1970; Bhattacharya & Giosan, 2003). Redistribution of sediment by longshore drift is an important process. Well-developed beaches can form down drift of the river mouth or updrift if longshore sediment transport is strong (Bhattacharya & Giosan, 2003).

Tide-dominated deltas form in areas affected by large tidal ranges (Galloway, 1975). Tidal currents moving in and out of the river mouth rework the sand into tidal bars, islands, and inlets. Deltas of this type are dominated by mud and sand without evidence of wave reworking. As documented below, most of the New Jersey deltas are mixed-energy, wave and tide dominated (Kulpecz et al., 2008).

Delta front facies from the modern Niger Delta (Allen, 1970) are characterized by very fine sand and clean silt that contain mica and finely comminuted plant debris. Prodelta facies from the Ba Lat Delta (a mixed river and wave dominated delta; van den Bergh *et al.*, 2007) consist of muddy sediments with quartz, feldspar and mica in the sand fraction and frequently display parallel laminations. Pyrite is sometimes prominently present in the sediments. Similar facies are observed in our coreholes as outlined below.

Estuarine environments are difficult to differentiate from lagoonal bay environments, because both display mixtures of fluvial and marine influences (Dalrymple *et al.*, 1992). Channel gravel deposits in the absence of extensive marine facies can be used to recognize estuarine versus lagoonal/tidal delta environments though these criteria are often missed.

Delta front-prodelta and estuarine observations

Examples of delta-front sediments are found in most boreholes on the New Jersey coastal plain (Island Beach, Atlantic City, Cape May, Bass River, Ancora, Ocean View, Millville, and possibly Fort Mott; Fig. 1), but they are largely absent from the Delaware coastal plain. Delta-front deposits are found in sequences with the most direct deltaic influence including the Miocene Kirkwood sequences, and the Cretaceous Cheesequake, Magothy, Bass River, and upper Englishtown sequences. Sediments, typically, are laminated with few cross-beds and uncommon bioturbation. Facies from the upper Englishtown sequence in the Sea Girt corehole consist of fine- to medium-grained, micaceous sands with thin beds (3–6 cm) and laminae of sandy clayey silt, clay, and lignite. Lignite occurs in laminae and in disseminated form, and it is more abundant in thicker beds. The sands are bioturbated and contain scattered clay-lined burrows (?*Ophiomorpha*). There are no obvious channel structures and no thick organic-rich beds. The presence of nannofossils indicates that these delta-front facies were marine. These sediments, which are similar to those described from the Niger Delta (Allen, 1970), are interpreted to represent delta-front environments.

Prodelta facies are found in association with delta front sediments. They consist of finely laminated, lignitic, micaceous clayey silts with scattered sand laminae and scattered pyrite nodules. The dominance of laminae with common lignite and mica, similar to prodelta sediments from Ba Lat Delta, suggest that these sediments were deposited in a prodelta environment (Owens & Gohn, 1985). In general, these deltaic facies coarsen upwards from the finer grained prodelta facies below to coarser grained delta-front facies above. The contact between the two is usually quite sharp with sand directly on top of laminated clay. Calcareous fossils are generally rare in prodelta facies, in part, because the high organic content is believed to make the ground water slightly acidic, dissolving the carbonate. Diatoms can be common in the absence of carbonate.

Cretaceous deltaic sediments in New Jersey (Kulpecz *et al.*, 2008) display distinct facies characteristics produced by varying degrees of wave and tide influence (Fig. 7). Kulpecz *et al.* (2008) noted that most New Jersey delta lithofacies exhibit the characteristics of a wave-dominated system with some mixed tide-wave dominated systems. For example, Magothy sequences in outcrop display evidence consistent with a stronger tidal influence on the delta deposits (e.g. modern Niger delta). This includes flaser and wavy beds and rare bidirectional cross bedding with

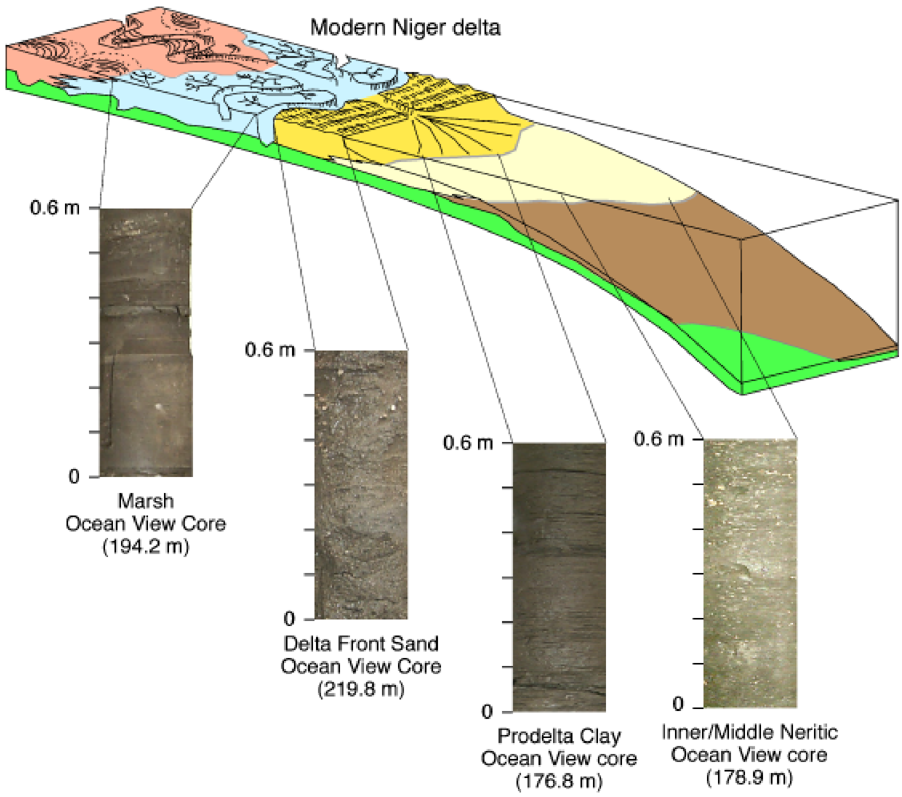

Fig. 7. Facies model for delta sedimentation (figure for the modern Niger River modified from Allen, 1970). Pictures correspond to the three major facies observed in New Jersey deltas; the corresponding location on the modern facies model is indicated.

clay drapes (Kulpecz *et al*., 2008). However, we note the scarcity of facies attributed to lower delta plain environments in our cores is consistent with a limited tidal influence on the coastal plain.

Facies interpreted to represent estuarine environments, seaward portions of drowned river valleys (Dalrymple *et al*., 1992), were rare and thin. They were recognized by the association of tidal flats and lagoonal clays with fluvial sands and gravels.

Shoreface to shelf criteria

Wave-dominated shoreline facies were distinguished from delta-front/prodelta facies in that they contain less organic matter, and more intense bioturbation. We have been unable to distinguish hummocky and swaley cross-stratification sets in the limited view afforded by the cores. Subenvironments recognized in a wave-dominated shoreline include the foreshore, proximal part of the upper shoreface, distal part of the upper shoreface, and lower shoreface (Harms *et al*., 1975, 1982; McCubbin, 1982; Fig. 8). Foreshore and proximal upper shoreface sediments consist of fine to coarse, well-sorted sand, with opaque heavy mineral laminae highlighting cross bedding (e.g. Harms *et al*., 1975, 1982; McCubbin, 1982). Distal upper shoreface sediments consist of fine to medium sand with admixed silts and less common clay layers (Harms *et al*., 1982, 1995; McCubbin, 1982). Laminae and rare cross beds are sometimes preserved but physical structures tend to be obscured by moderate to heavy bioturbation. Lower shoreface sediments consist of interbedded fine and very fine sands, commonly silty due to mixing, and commonly very shelly with whole shells preserved (Harms *et al*., 1975, 1982; McCubbin, 1982). These sediments are deposited below fair-weather wave base but above storm wave base and are commonly heavily bioturbated (Fig. 8).

Offshore environments (middle to outer neritic; $> \sim 30$ m water depth) consist of bioturbated silts and clays with rare thin sand beds (Fig. 8). Facies changes in these sections are subtle and can rarely be used to interpret water depth variations. In such environments, water depths are interpreted using benthic foraminifers. Sedimentary successions in a normal-marine shelf setting deposited below storm wave base consist of clay and silt with minimal quartz sand.

Shoreface to shelf observations

The strong deltaic influence seen in Miocene sections of New Jersey is less conspicuous in the Miocene of Delaware (Fig. 8). In regions and time intervals without deltaic influence, nearshore sediments accumulated in storm-dominated shoreface and neritic environments (e.g. Harms *et al*., 1975, 1982; McCubbin, 1982). A typical sequence (i.e. sequence C3, 322–299 m) consists of fine-grained offshore silts and clays overlain by bioturbated shelly fine sand representing the lower shoreface with fine to medium sand and broken shells from an upper shoreface. HSTs from

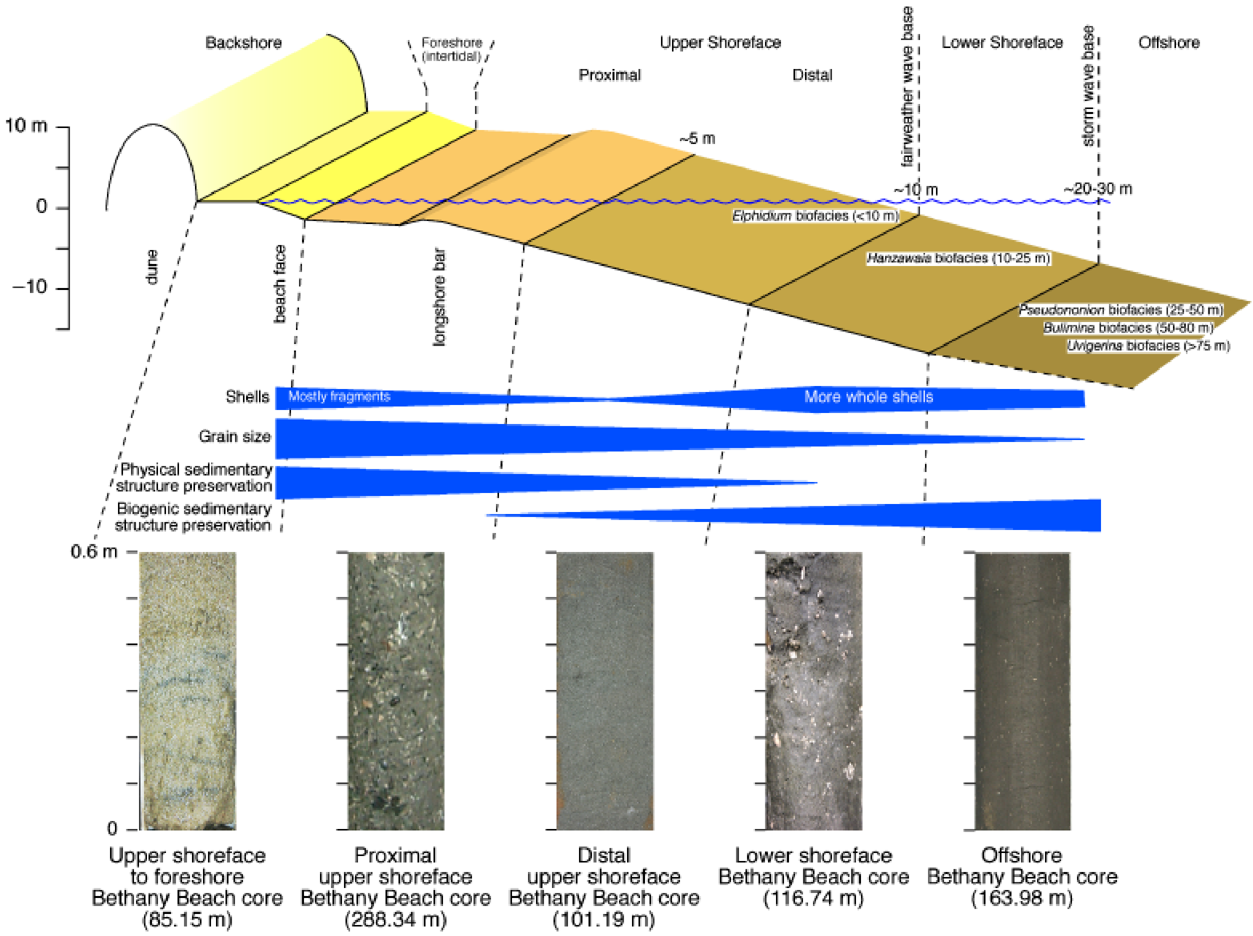

Fig. 8. Facies model for wave-dominated shoreline (Harms *et al.*, 1975, 1982; McCubbin, 1982). *Elphidium*, *Hanzawaia*, *Pseudononion*, *Bulimina*, and *Uvigerina* biofacies refers to benthic foraminiferal biofacies defined in Miller *et al.* (1997a, b). Core photographs are from the Bethany Beach borehole at the depths indicated below the photographs.

Delaware sequences contain facies deposited in progressively shallower marine environments showing the regressive character of the sections. In New Jersey, above a sequence boundary the TST consists of glauconitic clay to very clayey, glauconite sand. Just above the TST, clay and carbonate in the form of foraminifer tests are at a maximum for the sequence; this represents the MFS separating the TST from the HST. The HST coarsens upwards and at the top may contain a significant quartz sand fraction. Sediments throughout are generally heavily bioturbated.

DEPOSITIONAL PHASES

Mid-Atlantic region evolved through eight phases during the Early Cretaceous through the Miocene. Though facies vary along dip and strike due to expected changes in depositional environments, we note that there are periods (millions of years) when facies in our study area conform to specific depositional systems. General trends include: fluvial to deltaic sedimentation from 120–98 Ma and 92–85 Ma; delta and delta influenced environments from 98–92 Ma, 85–71.5 Ma and 24–12 Ma; a starved marine ramp from 71.5 to 42 Ma, a starved siliciclastic shelf from 42 to 24 Ma; and a low accommodation, eroded coastal system from 12 Ma to the recent.

Initial phase: fluvial environments (Barremian–Late Cenomanian; 120–98 Ma)

The oldest sediments (Barremian to earliest Cenomanian age, ∼120–98 Ma) identified in the study area are assigned to the Potomac Formation (pollen Zones I–III; Doyle & Robins, 1977; Table 1). The Ancora, Millville, and Sea Girt coreholes (Fig. 1) ended upon encountering the Potomac Formation. Only the Fort Mott corehole targeted Potomac sediments. Although basement was not encountered at Fort Mott, on the basis of gamma logs from nearby water wells the bottom of the corehole must have been just above basement, so a nearly complete section of the Potomac Formation was recovered (Sugarman *et al.*, 2004). We tentatively identify three sequences (Potomac 1–3) in the Potomac Formation based on studies at Fort Mott (Fig. 6; Sugarman *et al.*, 2004). The Potomac Formation was deposited during the first stages of thermo-flexural subsidence in the coastal plain when accommodation rates were highest (Watts & Steckler, 1979; Kominz *et al.*, this volume), yet sediment supply was plentiful enough to fill the basin.

Most of the Potomac Formation (Table 1) from the Fort Mott corehole is interpreted to represent anastomosed river environments in an upper delta plain (Sugarman *et al.*, 2004; Figs 1, 5 and 6). The degree to which marine processes influenced deposition of the Potomac sediments remains uncertain. Fine-to-coarse quartz sand that is locally lignitic and micaceous is generally interpreted to represent channel sands within the anastomosing environment. However, a few thick sand bodies (generally 15–30 m; representing a small percentage of the total thickness of the Potomac) appear to be regionally correlatable on gamma logs and could represent nearshore marginal-marine or delta-front environments. These sands lack the carbonate fossils (mollusks and foraminifers) that would be expected in a marine environment, but they also contain little evidence of channels in the form of fining upward successions that might be expected of a fluvial environment.

Second phase: marine systems with deltaic influence (Cenomanian–Early Turonian; 98–92 Ma)

Following the initial stage of deposition, a regional unconformity (Owens & Gohn, 1985) separates the Potomac Formation from the Raritan and Bass River Formations (Cenomanian–lower Turonian; Table 1). This unconformity represents a shift from predominantly terrestrial sediments to marine sediments but with evidence for the nearby presence of a delta. Three sequences have been identified in the Bass River Formation (Fig. 9) that correlate through multiple coreholes. An additional two older Bass River sequences are known only from the Sea Girt corehole, and their regional significance is not certain. For instance, they might be sequences (i.e. bracketed by regionally significant unconformities) or parasequences (i.e. bracketed by local or regional flooding surfaces). To the north at Sea Girt, Bass River sequences contain marine sediments that appear to have been deposited under deltaic influence, in agreement with findings that there was a deltaic source to the north at this time (Kulpecz *et al.*, 2008).

The Sea Girt borehole recovered Bass River sequences further to the north than any other New Jersey coastal plain core. Muscovite is common in Bass River sequences at Sea Girt; chlorite, typical of the formation to the south at Bass River, Ancora, and Millville (Miller *et al.*, 2006), is less common. The two basal sequences (Bass River 0 and Bass River 0.5), known only from Sea Girt, are coarsening upward successions deposited in delta-plain and prodelta-delta front environments, respectively. A major mid-Cenomanian hiatus associated with a large, rapid eustatic drop occurred between deposition of the BR0/0.5 and BR1 sequences and the facies above the associated unconformities are distinct from the facies below.

The Bass River I sequence is found in the Millville, Ancora and Sea Girt coreholes (Fig. 1). To the north at Sea Girt, it consists of prodelta and delta-front facies (Fig. 9). At Ancora, it consists of lower shoreface and possibly offshore deposits overlain by prodelta silt. To the south at Millville, it was deposited in lower shoreface and middle neritic environments with little evidence for deltaic influence. The Bass River II sequence (Fig. 9) appears to be middle neritic and lower shoreface shelly, burrowed sandy silt in the Millville and Ancora coreholes. At Sea Girt, prodelta sandy silt at the base of the sequence is replaced upsection by heavily burrowed sandy silt and clay. It is not clear from this core data where the main delta was at this time.

The Bass River III sequence consists of inner to middle neritic sand and clay. It generally represents deeper water deposition than other Bass River sequences. At both the Bass River and Sea Girt coreholes, the Bass River III sequence is relatively thick [> 140 ft (42.7 m) in the Bass River corehole], and there is evidence for five flooding surfaces (Sugarman *et al.*, 1999; Miller *et al.*, 2006). Ocean anoxic event two is interpreted to be preserved in the basal part of this sequence (Sugarman *et al.*, 1999). The Bass River III sequence at Bass River illustrates the sediments of a typical middle to outer neritic sequence. Above a MFS bounding a glauconitic interval, planktonic foraminifera indicate an upsection shallowing from middle to inner neritic environments with abundant epistominids. Coarse fraction data indicate further shallowing with increasing concentrations of quartz sand and mica and the presence of storm shell beds. Upsection, increased shelly sand and fewer clay beds indicate further shallowing to neritic and delta front environments. The sequence boundary at the base of the sequence was not penetrated at the Bass River corehole but the presence of glauconite at the base of the core is evidence that the sequence boundary was not far below.

Third phase: terrestrial and delta system (Late Turonian–Coniacian; 92–85 Ma)

A major mid-Turonian eustatic lowering is associated with the sequence boundary between BRIII and the Magothy sequences (Fig. 9). Magothy sequences were sampled in five coreholes: Bass River, Ancora, Fort Mott, Millville, and Sea Girt (Fig. 1). Middle neritic sediments that dominated the Bass River sequences were replaced by sediments representing upper delta plain, lower delta plain and delta front environments during Magothy time. In general, Magothy sediments are more terrestrial (upper delta plain) to the south in the Millville corehole and more marine (lower delta plain) to the north in the Sea Girt corehole.

The Magothy Formation (Table 1) consists of three to five sequences in the study area. The oldest sequence (MI) is found only at Bass River and Sea Girt (Pollen Zone IV). At Bass River, the sequence consists of prodelta clay that coarsens upwards to a delta-front sand. At Sea Girt,

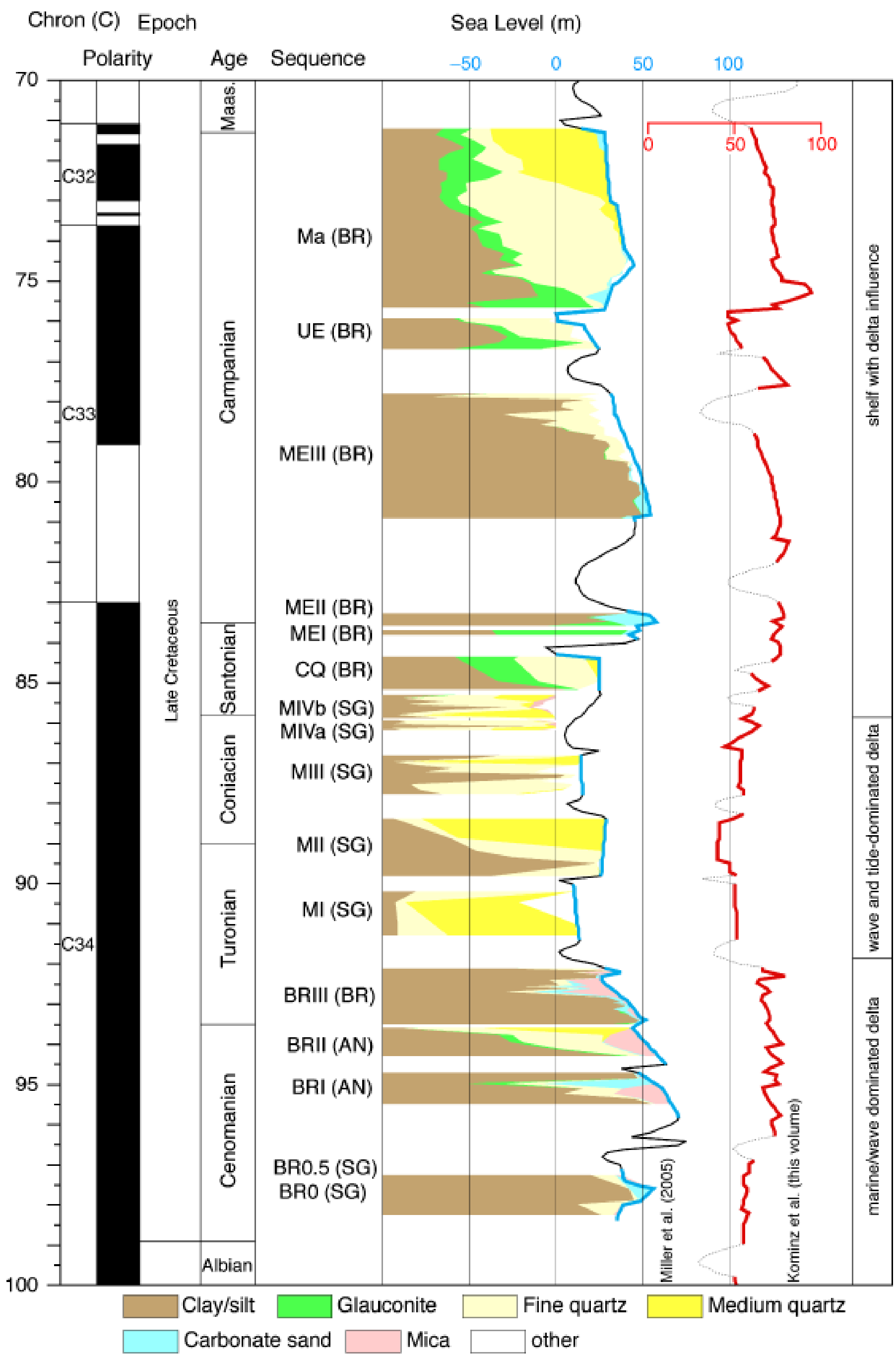

Fig. 9. Sequences and sea level, 70–100 Ma. Sequence notation is defined in Miller *et al.* (2004). Red curve – sea-level record from Kominz *et al.* (this volume); blue curve – sea-level record from Miller *et al.* (2005) for comparison.

the MI sequence facies are interpreted as estuarine and deepen upsection possibly indicating that the HST was not preserved.

The second sequence (MII) consists of upper-delta-plain sediments in the Millville, Ancora and Sea Girt coreholes. The third sequence (MIII) represents fluvial channels prograding over estuarine sediments in the Fort Mott, Ancora and Bass River coreholes. To the north in Sea Girt, the sediments represent a lower delta-plain environment.

Sequences IVa and IVb are found only in the Sea Girt corehole (Miller *et al.*, 2006; Kulpecz *et al.*, 2008) and require regional verification. Sequence IVa (older) represents tidal channels alternating with marsh deposits. Sequence IVb (the youngest Magothy sequence) represents a bay or lagoon overlain by upper shoreface delta front sands.

In general, the sequences in the Magothy represent an initial dramatic shallowing of water depth on the New Jersey coastal plain followed by sequences that become progressively less terrestrial and more marine through time. The Magothy sequences are bracketed by two major sequence boundaries (mid-Turonian and Coniacian–early Santonian) and represent a general period of lower long-term sea level.

Table 1. Formations found in the New Jersey coastal plain

Age	Formation
Cenozoic	
Miocene & younger	Cape May
	Cohansey
	Kirkwood
Oligocene	Sewell Point
	Atlantic City
Eocene	Absecon Inlet
	Shark River
	Manasquan
Paleocene	Vincentown
	Hornerstown
Upper Cretaceous	
Maastrichtian	Red Bank/Navesink
Santonian & Campanian	Mt. Laurel
	Wenonah
	Marshalltown
	Englishtown
	Woodbury
	Merchantville
	Cheesquake
Cenomanian to Turonian	Magothy
	Bass River/Raritan
Lower Cretaceous	
Barremian to Albian	Potomac
Paleozoic	
Paleozoic	Basement

Fourth phase: marine with a strong wave-dominated delta influence (Santonian–Campanian; 85–71.5 Ma)

Marine conditions overprinted by a delta on the New Jersey coastal plain returned during the Santonian and continued through the Campanian (Fig. 9). This switch is believed to represent increased accommodation due to relative sea-level rise (see 'Discussion').

The Cheesequake Formation and sequence (Table 1) is a poorly sampled thin sequence. There is no discernable pattern to the distribution of its sediments. The Cheesequake sequence is between the terrestrial Magothy sediments below and more deltaic sediments above. In outcrop, the formation is a silt (Litwin *et al.*, 1993). Similar lithologies are found at Sea Girt (very poor recovery) and Fort Mott, where they are interpreted as middle neritic deposits. At Bass River and Ancora, correlative sediments are marine sand and at Millville the section is interpreted as a delta front deposit.

The overlying Merchantville, Englishtown, and Marshalltown sequences all show the effects of a large delta on marine sedimentation. Two, thin glauconite-rich sequences (MEI, MEII; Santonian to lower Campanian) are concatenated at the base of the Merchantville Formation (Table 1). These two sequences are well expressed lithologically only in the Sea Girt corehole. There, inner-middle neritic glauconite sand (TST) grades up to inner neritic sand with delta influence (HST). To the south, HSTs are absent, and it is very difficult to distinguish sequence boundaries separating the TSTs. MEIII is a thick lower Campanian sequence encompassing the upper Merchantville, Woodbury and lower Englishtown Formations (Table 1). The sequence is generally thin to the north, where it consists of prodelta and delta front facies, and thicker with classic prodelta facies to the south. The upper Englishtown (mid-Campanian) and Marshalltown (upper Campanian) sequences are both thicker with more deltaic facies in the north at Sea Girt than to the south where they display more marine facies.

Low sedimentation rates are indicated by the glauconitic sediments found in the Merchantville Formation (Sugarman *et al.*, 1995). The first significant occurrence of quartz sand is at the top of the MEIII sequence in the lower Englishtown Formation. The upper Englishtown sequence has more quartz, but is only thick in the north at Sea Girt. The Mount Laurel Formation at the top of the Marshalltown sequence is the first thick, widespread marine sand on the New Jersey coastal plain and must represent an increase in sediment input. The New Jersey coastal plain water depths generally shallow through this interval from the Santonian to late Campanian.

Fifth phase: starved marine ramp (Maastrichtian–middle Eocene; ~71.5–42 Ma)

Sea level stood very high during the Maastrichtian through the middle Eocene (71.5–42 Ma; Fig. 10), and most sediments were deposited in middle (30–100 m) to outer neritic (> 100 m) paleoenvironments. Deeper environments (i.e. bathyal) are not represented in onshore Cretaceous coreholes. Sediments were deposited on a ramp-style margin (Steckler *et al.*, 1999). There was an uninterrupted gradient on the continental shelf from the inner neritic zone to the deep sea with, apparently, no sharp continental shelf break offshore of New Jersey. The gradient seaward of the basement hinge line was 1 : 500. The gradient landward of the hinge line (which lies immediately offshore today) was probably similar to the gradient on the modern shelf (1 : 1000; Steckler *et al.*, 1999).

The Maastrichtian and lower Paleocene are marked by glauconite, carbonate, and clay deposition. Sequences are typically difficult to distinguish in the cores of the absence of siliciclastic sediment in the highstands. The sequences consist of packages of glauconitic sand (TSTs) overlain by glauconitic clays (typically carbonate rich; HSTs) separated by unconformities. The influx of clay is interpreted to represent a very small amount of progradation during highstand. The sediments are burrowed and sequence boundaries are difficult to identify. Evidence of the latest Cretaceous impact and extinction is well preserved in some of the coreholes as spherules and a reworked Cretaceous clast layer (Olsson *et al.*, 1997). In other cores the impact layer has been destroyed due to erosion and bioturbation, and distinguishing the Cretaceous from the Paleocene on lithologic grounds is difficult.

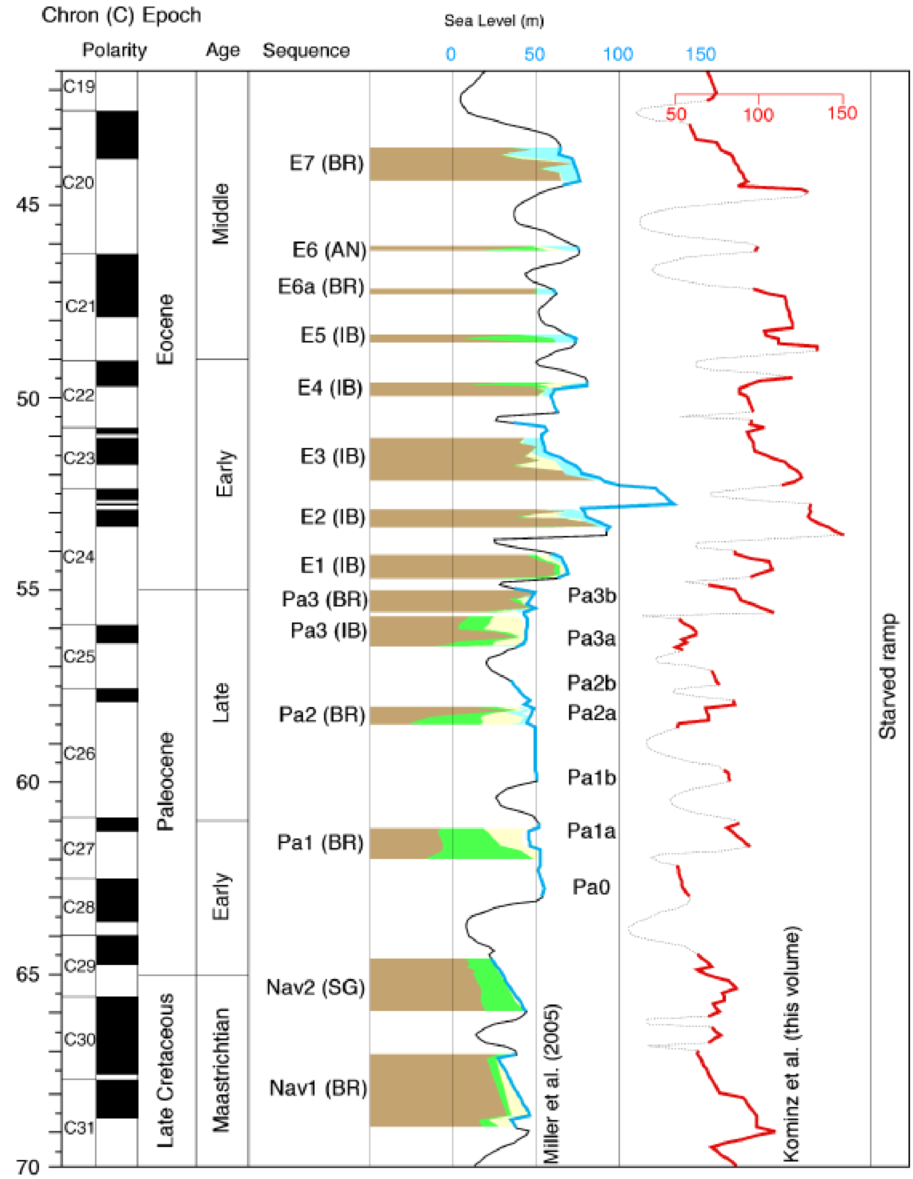

Fig. 10. Sequences and sea level, 40–70 Ma. E sequence notation is defined in Browning *et al.* (1997). Pa sequence notation is defined in Liu *et al.* (1997). Nav sequence notation is defined in Miller *et al.* (2004). BR, Bass River core; AN, Ancora core; IB, Island Beach core; SG, Sea Girt core. Red curve – sea-level record from Kominz *et al.* (this volume); blue curve – sea-level record from Miller *et al.* (2005) for comparison. See Fig. 9 for lithology key.

Upper Paleocene sequences again resemble typical New Jersey sequences indicating an increase in the amount of sediment coming onto the shelf. The sediments are probably delta-influenced as shown by the abundant organic material and laminations in the cores. Sediments from the uppermost Paleocene contain evidence for the Paleocene–Eocene thermal maximum (Cramer *et al.*, 1999). These clays contain little sand or silt, and foraminfer tests show the oxygen and carbon isotopic shifts associated with the event.

Deepest water depths of the Cenozoic, shown by foraminiferal biofacies, were attained during the early Eocene. Water depths, determined using benthic foraminiferal biofacies (Browning *et al.*, 1997), show middle to outer neritic paleodepths. Sequences consist of thin glauconitic clays overlain by carbonate-rich clays with layers of porcellanite interspersed. Sequence boundaries are inferred by matching the few surfaces in the cores with biostratigraphic breaks. Above the lower/middle Eocene boundary, there are three, thin, glauconite-rich sequences. The lower-middle Eocene, although still carbonate rich, is slightly shallower (middle neritic), and sequences can be distinguished lithologically particularly by basal glauconite sands.

Sixth phase: starved siliciclastic shelf (late Middle Eocene–Oligocene; 42–24 Ma)

The transition to the ice-house world began in the late middle Eocene (Miller *et al.*, 1991) and was associated with

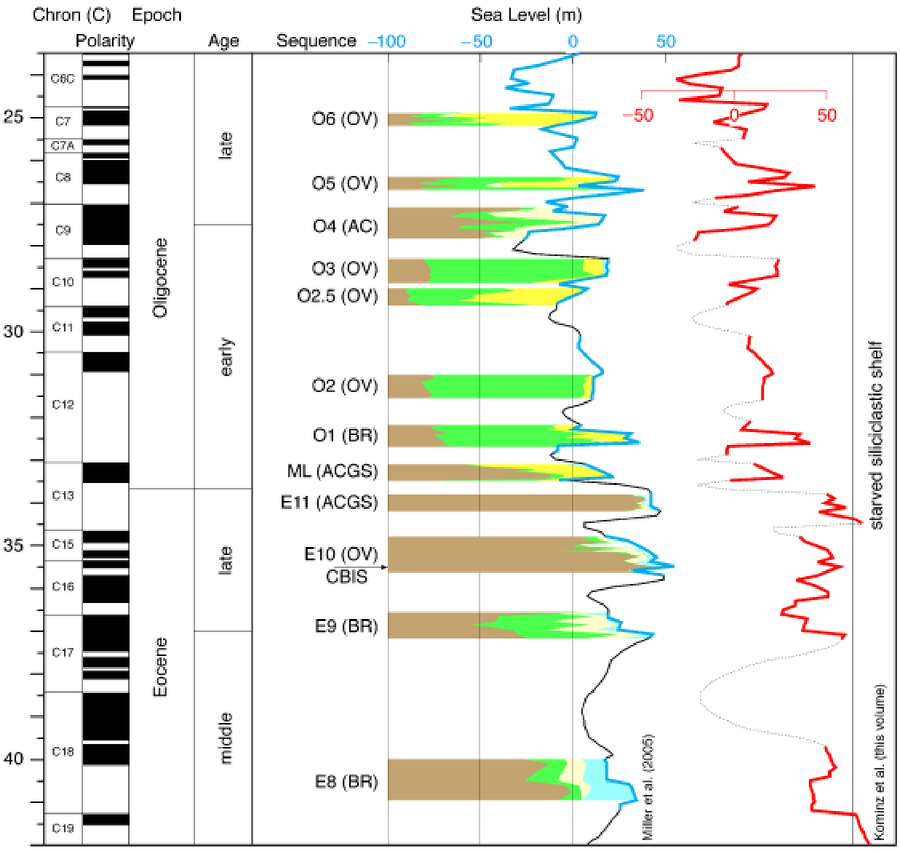

Fig. 11. Sequences and sea level, 24–42 Ma. O and ML sequence notations are defined in Pekar *et al.* (2000). E sequence notation is defined in Browning *et al.* (1997). OV, Ocean View core; BR, Bass River core; CBIS, Chesapeake Bay impact structure. Red curve – sea-level record from Kominz *et al.* (this volume); blue curve – sea-level record from Miller *et al.* (2005) for comparison. See Fig. 9 for lithology key.

a new phase of deposition on the mid Atlantic margin (Fig. 11). Sediments on the shelf shifted from carbonate-rich clays to siliciclastics beginning onshore in the late middle Eocene, and culminating on the modern continental slope at the Eocene/Oligocene boundary (Miller *et al.*, 1998a, b). The sediments also changed the nature of the shelf from a carbonate ramp to a prograding siliciclastic margin that fully developed by the latest Oligocene (~27 Ma pulse; Miller *et al.*, 1997a, b) to early Miocene.

An abrupt shift from carbonate marls to siliciclastic sediments takes place across the study area at approximately 42 Ma (Browning *et al.*, 1996). This shift is interpreted as a response of the margin to an expansion of ice on Antarctica and a lowering of global sea level (Browning *et al.*, 1996). High sea level returned briefly in the late Eocene, and most of the coreholes record a fine-grained middle-outer neritic clay. The Eocene/Oligocene boundary is typically a dramatic burrowed contact with Oligocene glauconite sand overlying upper Eocene clay.

Oligocene sequences consist of heavily bioturbated, glauconitic, quartz sand and subordinate glauconite clays. (Pekar *et al.*, 2000). Glauconite found in the HSTs is interpreted as recycled from older deposits rather than forming in place. There is little change in facies through the HSTs, and sequence boundaries are difficult to delineate because they consist of glauconite clayey sand overlying glauconite sand. Age breaks, determined from Sr-isotopic stratigraphy, are an important tool used to identify likely sequence boundaries, which can then be confirmed lithologically. Lower Oligocene sequences are generally finer grained than upper Oligocene sequences. Input of quartz sand resumed in the latest Oligocene (~27 Ma; Miller *et al.*, 2005).

Seventh phase: marine shelf with wave-dominated delta influence (early and middle Miocene; 24–12 Ma)

The lower-middle Miocene Kirkwood Formation (Table 1) predominantly consists of packages of delta-front sands overlying prodelta clays (Fig. 12). These packages reflect the strong influence of one or more riverine sources of sediments in New Jersey at the time. The oldest Kw0 sequence is only thick at Cape May where it contains abundant glauconite sand.

The Kwla sequence is one of the most widespread of all the Miocene sequences. It is very thin to the north in Sea Girt, and the delta facies are less well developed in Cape May. It is thickest at Ocean View, and the delta facies are best developed in the cores from the central part of the New Jersey coastal plain. This indicates a central coastal plain source at this time. Shallow marine sand at Cape

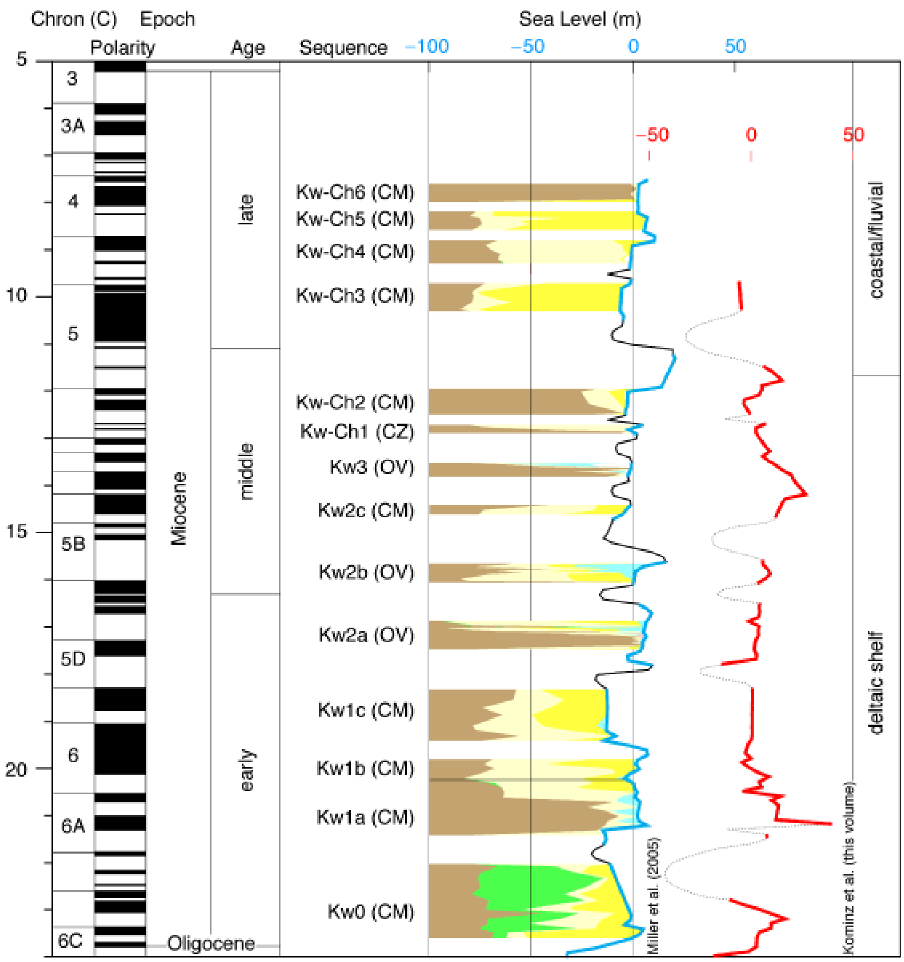

Fig. 12. Sequences and sea level, 5–24 Ma. Kw sequence notation refers to sequences defined in Miller *et al.* (1997a, b). CM, Cape May core; OV, Ocean View core; CZ, Cape May Zoo core. Red curve – sea-level record from Kominz *et al.* (this volume); blue curve – sea-level record from Miller *et al.* (2005) for comparison. See Fig. 9 for lithology key.

May fits the wave-dominated delta model with longshore currents transporting sediment from north to south.

The Kwlb is a thin but distinct sequence, that has no discernable time gap separating it from the Kwla sequence. It is best developed in cores in the central part of the state (Island Beach, Bass River, Ancora, and Atlantic City; Fig. 1). To the north and south, the deltaic sediments are interbedded with neritic shelf sediments. Lower shoreface sands are best developed to the south at Cape May and Cape May Zoo coreholes. To the north, Sea Girt is dominated by bay and lagoonal sediments.

The Kwlc sequence is only found in the most down-dip cores (Cape May and Cape May Zoo sites) in the Cape May Peninsula. The sequence is presumed to have been eroded from the more updip sites. Most of the sediments in the sequence were deposited in middle neritic and inner neritic (lower shoreface) environments, and there is little evidence for the presence of a delta.

The Kw2a sequence is represented by well-developed prodelta clays. The sand facies are not lignitic or micaceous and are interpreted as lower shoreface and upper shoreface deposits that do not resemble typical delta-front deposits. The sequence is well-dated only at the Island Beach, Atlantic City, Cape May, Ocean View, and Cape May Zoo coreholes (Fig. 1). It is interpreted to be present at Bass River and Island Beach. Thin, undated sequences at Millville and Sea Girt cannot be confidently correlated to the Kw2a sequence. The absence of delta-front sands in the presence of prodelta clays is puzzling. It may be that very high wave energy at this time winnowed the lignite and mica from the nearshore deposits.

The Kw2b sequence preserves evidence for a small delta situated between the Ocean View and Cape May Zoo coreholes. Only the Cape May Zoo corehole has prodelta sediments. Other cores containing the sequence record nearshore and inner neritic sediments that are not delta influenced. The Kw2c sequence (only found at Cape May) is inner neritic with a thin prodelta recorded at top. The Kw3 sequence at Ocean View and Atlantic City has prodelta and delta-front sediments. At Cape May and Cape May Zoo, there are few delta sediments indicating the river source was located between Ocean View and Atlantic City.

The two Kirkwood-Cohansey (Kw-Ch) sequences are poorly dated and therefore difficult to correlate. Sediments assigned to the Kw-Ch sequences are found only in the four most downdip coreholes (Cape May, Cape May Zoo, Ocean View and Atlantic City; Fig. 1). These thin sequences generally consist of prodelta clays overlain by delta front and lower shoreface sands. By this time, the main depocenters were offshore and thin TST/HST packages accumulated landward of the depocenter in an area of limited accommodation. The area represented by the New Jersey coastal plain was largely a zone of sediment bypass.

Eight phase: low accommodation, eroded coastal system (late Miocene-Recent; 12–0 Ma)

The Cohansey, Stone Harbor, and Cape May Formations (Table 1) represent sedimentation of the past 10 Myr (Fig. 12). The Cohansey and Stone Harbor Formations are laterally adjacent formations (Newell *et al.*, 2000) but it is not certain they are time equivalent. The Stone Harbor Formation is an estuarine deposit found in the Cape May Peninsula (Sugarman *et al.*, 2007). It consists of interbedded organic-rich medium to coarse quartz sand, fine micaceous quartz sand, and laminated sandy clay. These interbedded facies represent a variety of estuarine subenvironments, including channels, fringing marshes and bays. To the north, the Cohansey Formation is a 'yellow' sand with minor clay that represents barrier island and fluvial environments. The Cape May Formation consists of Pleistocene and Holocene nearshore deposits representing recent marine transgressions. There does not appear to be any sediment deposited between the youngest Cohansey Formation ($\sim$7 Ma) and the Pleistocene, though, poorly dated upland fluvial gravels (Pennsauken, Beacon Hill; Pazzaglia & Gardner, 1994) may be upper Miocene–Pliocene. These gravels form the highest peaks in the coastal plain.

DISCUSSION

Million year-scale sequences

Prior studies of outcrops and cores from the New Jersey and Delaware coastal plains documented Myr-scale (equivalent to EPR third- and second-order sequences; Vail *et al.* 1977; Haq *et al.*, 1987) sequences that correlate across the region (Fig. 2) and with other basins (e.g., Miller *et al.*, 1996a, b, 1998a, 2004, 2005; Pekar & Miller, 1996). There is no distinct order of our sequences that show a weak periodicity of $\sim$3 Myr (Miller *et al.*, 2005) though we have previously linked the sequences to both so-called second- and third-order sequences (Vail *et al.*, 1977; Haq *et al.*, 1987). 'Icehouse' (Oligocene and younger) sequences can be linked firmly to deep-sea $\delta^{18}O$ increases that are a proxy for glacioeustatic sea-level falls (Miller *et al.*, 1996a, b, 2005) and are thus global in extent. Several 'Greenhouse' (Late Cretaceous to Eocene) sequences can similarly be linked to deep-sea $\delta^{18}O$ increases, leading Miller *et al.* (2003a, 2004, 2005) to postulate that small, ephemeral ice sheets existed during 'cool snaps' in the 'Greenhouse' world (Royer *et al.*, 2004). This study adds several new sites to these comparisons (Sea Girt, Millville, Fort Mott, Cape May Zoo; Fig. 2). For the Late Cretaceous, the first three new sites are critical because previous studies relied solely on Bass River and Ancora (Miller *et al.*, 2003a, 2004). The recent Cape May Zoo site also validates several Miocene sequences previously recorded only at Cape May and Bethany Beach (Fig. 2). The within-sequence variations in these sequences are well established (Fig. 3) and provide a predictable succession of aquifer sands and silty-clay confining units (Sugarman & Miller, 1997; Sugarman *et al.*, 2005a).

10-Million year-scale sequences

Although each sequence is unique and may contain a variety of depositional environments, in the study area we find thick sections (representing long time periods) under the modern coastal plain that are typified by similar depositional styles. Overlap is expected, but it is clear that some periods were dominated by deltaic vs. nondeltaic and high vs. low sedimentation rate. These thick sections are referred to here as depositional phases.

As developed in detail above, we show that the study area on the mid-Atlantic coastal plain evolved through eight different depositional phases (Figs 2, 9–12). The Delaware record is spottily sampled except for one Miocene corehole (Bethany Beach) and most of the comparisons discussed below necessarily focus on the New Jersey record. We compare these stages to the longer-term ($\sim$10 Myr) sea-level changes (Figs 2, 9–12) to evaluate the effects of long-term sea-level changes on the margin. Facies evolution through these depositional phases was controlled by changes in accommodation rate, long-term eustatic changes, and variations in sediment supply. The facies themselves clearly show the influence of changes in sediment source, with deltas prograding through the study area. Detailed mapping using well logs in addition to continuous cores has shown that river systems were apparently absent in New Jersey during portions of the Late Cretaceous (Kulpecz *et al.*, 2008). At other times well log correlations indicate one to two depocenters showing river influence (Kulpecz *et al.*, 2008). These river systems were largely absent in the Paleogene, and returned during the latest Oligocene–Miocene (27–12 Ma).

The initial phase (Barremian–earliest Cenomanian; $\sim$120–98 Ma) consisted of primarily alluvial plain, anastomosing riverine environments of the Potomac Formation (Fig. 9). Though uppermost Jurassic to lower Lower Cretaceous deposits are found in the Atlantic coastal plain to the south and offshore (Owens & Gohn, 1985), the New Jersey and Delaware coastal plains formed some time after rifting ($\sim$180 Ma). Deposition began in New Jersey approximately in the Barremian (i.e. slightly older than 120 Ma), though deposition in Delaware may predate this (Olsson *et al.*, 1988). This earliest phase of deposition was dominated by high sediment supply because it was associated with: (1) very rapid thermoflexural subsidence and thus high accommodation rates (Kominz *et al.*, 1998; this volume); (2) high accommodation rates filled with fluvial sediments that imply sediment supply was relatively high and the basin was filled to capacity (i.e. resulting in fluvial deposition); this is testified to by the great thickness and high accommodation rates of the Potomac Formation (e.g. in the Cape May region the Potomac Formation is 1 km thick out of 1.8 km total sediment thickness); (3) long-term sea level that was slightly lower than subsequent phases (e.g. +35 to 45 m vs. ~75 m in the next phase; Kominz *et al*., this volume); and (4) very warm climates as testified to by floral data (Wolfe & Upchurch, 1987) and global temperature proxies (Huber *et al*., 1995). These conditions resulted in megathermal, anastomosing riverine environments with extensive soil formation in our study area.

Neritic shelf systems with a minor deltaic influence occurred in the Cenomanian–Turonian (~98–92 Ma) during deposition of the Bass River sequences (Fig. 9). Our studies show that long-term sea level was relatively high (~75 m; Figs 2 and 9) and that there were moderate rates of accommodation. The Bass River sequences (BR0-3) contain prodelta sediments along with some delta front deposits, though deposition was primarily on an inner to middle neritic, storm-dominated shelf. The Bass River sequences were deposited during one of the two Late Cretaceous peaks in global warmth at ~93 Ma as indicated by deep-sea $\delta^{18}O$ records (Huber *et al*., 1995; Miller *et al*., 2005) and recorded a major ocean anoxic event (OAE2; Sugarman *et al*., 1999). Thermal subsidence rates had begun to decrease by Bass River time (Fig. 2 in Kominz *et al*., this volume), and thus the increase in water depth cannot be due to thermal subsidence. Thus, this first marine incursion of the New Jersey coastal plain is attributed to long-term rise in sea level. This is evidenced by backstripped estimates that show sea level deepened from 50 to 80 m over this period, though a decrease in sediment supply may have contributed to this long-term transgression.

Wave and tidal dominated delta systems were important in the late Turonian through Coniacian (~92–85 Ma; Figs 2 and 9) during deposition of the Magothy sequences. Kulpecz *et al*. (2008) presented evidence for both tide-wave and wave dominated deltaic systems during Magothy time. Although there is a difference in types of sediments, the accumulation rate in the Magothy Formation is similar to that of the Bass River sequences and, thus, the differences in sedimentary environment between the two cannot be attributed to an increase in sediment input (Miller *et al*., 2006). We attribute the change from marine in the Bass River sequences to nonmarine and marginal marine environments in Magothy sequences to long-term eustatic fall (~75 m during Bass River time to ~45 m during Magothy time) coupled with moderate rates of subsidence.

Marine conditions with a strong wave-dominated deltaic influence occurred during the Santonian–Campanian (~85–71.5 Ma). This interval represents classic onshore New Jersey transgressive-regressive shelf sequences with basal deep water glauconite TSTs and prodelta and delta front facies in the HST (e.g. Sugarman *et al*., 1995). These facies were developed in a period of long-term (~15 Myr; *ca*. 85–71.5 Ma) sea level high (50–100 m; Fig. 9). Once again, the difference between the previous nonmarine Magothy and the marine sequences appears to be attributable to long-term rise in sea level (Kominz, 1984).

A starved carbonate ramp environment with pulses of deltaic influence occurred during the Maastrichtian–middle Eocene (~71.5 to ~44 Ma) due to high long-term sea level and low sediment supply (Fig. 10). In general, the Maastrichtian–early Paleocene was remarkably starved of siliciclastic input so that glauconite sedimentation dominated. Limited deltaic influence occurred in the late Paleocene. The early Eocene was characterized by remarkably uniform calcareous clay deposition that continued into the middle Eocene. Sediment starvation must be due to low input from the hinterland; nevertheless, the first-order control on deposition from the Maastrichtian to middle Eocene appears to be the generally high long-term sea level that peaked in the early Eocene. The cause of low sediment input is not clear because the Appalachian Mountains were not flooded and the ancient beaches would still have been located within modern New Jersey in the north and eastern Pennsylvania to the south.

A starved siliciclastic shelf occurred during the latest middle Eocene–Oligocene (~41–24 Ma) associated with moderately high sea level and very low siliciclastic sediment supply (Fig. 11). The exception is a pulse of quartz sand during deposition of sequences E8 and E9 (~41–36). A reduction of pelagic carbonate deposition cannot be explained by dilution due to increased sedimentation, because, in general, sedimentation rates were very slow (the exception is the high clay sedimentation during sequences E10 and E11; ~35–34 Ma). We attribute the reduction in pelagic carbonate to cooling that began in the late middle Eocene (Miller *et al*., 1987, 2005; Zachos *et al*., 2001). Cooling shut down carbonate production in New Jersey (now coastal plain), but carbonate sedimentation in deeper water (now continental slope) did not cease until the earliest Oligocene global cooling and ice volume event (Oi-1; Miller *et al*., 1996a, b). Oligocene sedimentation was excruciatingly slow ($<<$ 1 cm kyr^{-1}), with thin clinoform sequences prograding across the shelf (Pekar *et al*., 2000). Medium to coarse quartz sand reappeared in the latest Oligocene (~27 Ma), signaling a change that progressively resulted in thicker and coarser prograding sequences in the early to middle Miocene.

Marine environments with strong wave-dominated delta influence occurred during the early and middle Miocene due to a major increase in sediment supply and accommodation (Fig. 12). This change to thick, prograding sequences reached the modern middle shelf in the middle–late Miocene (Poag & Sevon, 1989; Pazzaglia & Gardner, 1994; Steckler *et al*., 1999). This sediment pulse is a first-order feature not only of this margin, but in fact of many margins (Lavier *et al*., 2001). Interpretations for the cause of this change in sedimentation range from tectonic changes in the hinterland (e.g., Poag & Sevon, 1989; Pazzaglia, 1993) to global cooling (see Steckler *et al*., 1999, for a review). The fact that this change to thick prograding clinoforms is associated with a middle Miocene cooling (Steckler *et al*., 1999) and is found on other margins (Lavier *et al*., 2001) argues for a climatic control. The large sediment influx would also have contributed to local increase in accommodation due to flexural loading (Browning *et al*., 2006).

Low accommodation and eroded coastal systems occurred in the onshore coastal plain during the last 7 Myr (Fig. 12). These are a result of low long-term sea level and sediment bypass. Thick accumulations of upper Neogene sediments occur offshore (Miller & Mountain, 1994), though the generally poor representation of the Pliocene anywhere on this margin remains a mystery.

CONCLUSIONS

Continuous coring has revolutionized our understanding of the evolution of the passive continental margin of the mid-Atlantic US. Pioneering studies previously documented Cretaceous–Miocene sequences and related them to global sea-level changes (Olsson, 1991), but unraveling the relative importance of eustasy, tectonics, and sediment input was not possible until coring by ODP Legs 150X and 174AX provided material for detailed sampling and analysis, development of a firm chronology (better than 1 Myr resolution), and a relatively precise paleowater depth history. This well-dated paleodepth record was backstripped to provide a sea-level estimate by accounting for the effects of compaction, loading, and thermal subsidence (see Kominz *et al*., this volume for detailed error analysis of this procedure). The resultant sea-level estimate can be used to evaluate the effects of long-term sea-level change, and thus evaluate the role of tectonics and changes in sediment supply. The US mid-Atlantic margin evolved through eight depositional phases over the past ~110 Myr. Long-term change in sea level is a first-order control on the nature of the depositional phases, with terrestrial or erosional environments during intervals of low sea level (Barremian–early Cenomanian; late Turonian–Coniacian; late Miocene–Recent) and marine deposition when sea level was high. Superimposed on this is the influence of sediment input, with starved, pelagic sections during low input (Maastrichtian–Oligocene) and riverine influenced, thick, rapid deposition during high input (Barremian–Campanian, early–middle Miocene). The tectonic overprint is minimal (e.g., typically < 30 m; Browning *et al*., 2006) except for the long-term exponential trend of decreasing thermal subsidence. Studies to date document that an eustatic estimate can be derived from regions such as the mid-Atlantic US, though we emphasize that our results must be considered a preliminary, testable model to be compared with records from other margins.

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