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RESPONSE OF LATE CRETACEOUS MIGRATING DELTAIC FACIES SYSTEMS TO SEA LEVEL, TECTONICS, AND SEDIMENT SUPPLY CHANGES, NEW JERSEY COASTAL PLAIN, U.S.A.

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ABSTRACT: Paleogeographic, isopach, and deltaic lithofacies mapping of thirteen depositional sequences establish a 35 myr high resolution (> 1 Myr) record of Late Cretaceous wave- and tide-influenced deltaic sedimentation. We integrate sequences defined on the basis of lithologic, biostratigraphic, and Sr-isotope stratigraphy from cores with geophysical log data from 28 wells to further develop and extend methods and calibrations of well-log recognition of sequences and facies variations. This study reveals the northeastward migration of depocenters from the Cenomanian (ca. 98 Ma) through the earliest Danian (ca. 64 Ma) and documents five primary phases of paleodeltaic evolution in response to long-term eustatic changes, variations in sediment supply, the location of two long-lived fluvial axes, and thermoflexural basement subsidence: (1) Cenomanian–early Turonian deltaic facies exhibit marine and nonmarine facies and are concentrated in the central coastal plain; (2) high sediment rates, low sea level, and high accommodation rates in the northern coastal plain resulted in thick, marginal to nonmarine mixed-influenced deltaic facies during the Turonian–Coniacian; (3) comparatively low sediment rates and high long-term sea level in the Santonian resulted in a sediment-starved margin with low deltaic influence; (4) well-developed Campanian deltaic sequences expand to the north and exhibit wave reworking and longshore transport of sands; and (5) low sedimentation rates and high long-term sea level during the Maastrichtian resulted in the deposition of a sediment-starved glauconitic shelf. Our study illustrates the widely known variability of mixed-influence deltaic systems, but also documents the relative stability of deltaic facies systems on the 10^6–10^7 yr scale, with long periods of cyclically repeating systems tracts controlled by eustasy. Results from the Late Cretaceous further show that although eustasy provides the template for sequences globally, regional tectonics (rates of subsidence and accommodation), changes in sediment supply, proximity to sediment input, and flexural subsidence from depocenter loading determines the regional to local preservation and facies expression of sequences.

INTRODUCTION

Sequence stratigraphy, the use of unconformity-bounded units and their constituent facies to correlate sedimentary sequences on regional scales (Mitchum et al. 1977), has been a powerful tool in predicting the distribution of important economic resources such as hydrocarbon reservoirs (Vail et al. 1977) and groundwater aquifers (Sugarman and Miller 1997; Sugarman et al. 2006). The application of sequence stratigraphic principles revolutionized our understanding of the New Jersey Coastal Plain (e.g., Olsson 1991) and established the Mid-Atlantic Margin as a natural laboratory for examining the fundamental mechanisms that control deposition on passive margins (Miller et al. 1998a; Miller et al. 2005). Processes that control deposition on this, and other margins, include eustasy (Miller et al. 2005), changes in the prevailing tectonic regime (subsidence versus uplift) (e.g., Browning et al. 2006), and variations in sediment supply (Pitman and Golovchenko 1983; Poag and Sevon 1989; Reynolds et al. 1991).

The sequence stratigraphic method has also proven useful in examining the evolution of deltaic systems. Sequence boundaries and flooding surfaces (e.g., Galloway 1989) represent temporally significant surfaces that can be used to establish the paleogeographic history of a deltaic margin or chart the distribution of facies through time and space. These studies of deltaic systems range in scope from the robust data sets of the Gulf of Mexico (Galloway 1989; Galloway et al. 2000; Combellas-Bigott and Galloway 2006) and Western Canada (Plint 2003), to high-resolution outcrop studies of the Cretaceous Western Interior Seaway that evaluate the higher-order response of deltaic sequences, parasequences, and facies to forcing mechanisms (e.g., Bhattacharya and Walker 1991; Lee et al. 2007; Gani and Bhattacharya 2007; Davies et al. 2006). Although these studies have contributed greatly to our understanding of ancient deltaic systems, many face complications such as complex tectonic histories and difficulty establishing geochronologic control due to shallow and nonmarine facies. Because each margin offers a unique blend of eustatic, tectonic, and sediment supply controls, differentiating the sedimentary response to each can be very difficult.

While the New Jersey Coastal Plain does not afford the broad regional picture of the Gulf of Mexico, or the detail of outcrop-scale facies analyses, it offers a long-term (~ 35 Myr) record of eustatically forced Late Cretaceous deltaic sequences with high temporal resolution (> 1 Myr) from the work of the Ocean Drilling Program Leg 174AX (Miller et al. 1998b; Miller et al. 1999; Miller et al. 2003; Sugarman et al. 2006: summarized in Miller et al. 2004 and this study). Because deposition occurred on a passive margin dominated by consistent thermoflexural subsidence (Kominz et al. 1998; Miller et al. 2004), this study avoids many of the difficulties associated with tectonically active settings.

The main objectives of this study are to: (1) reconstruct the Late Cretaceous paleogeographic evolution of this deltaic margin; (2) examine

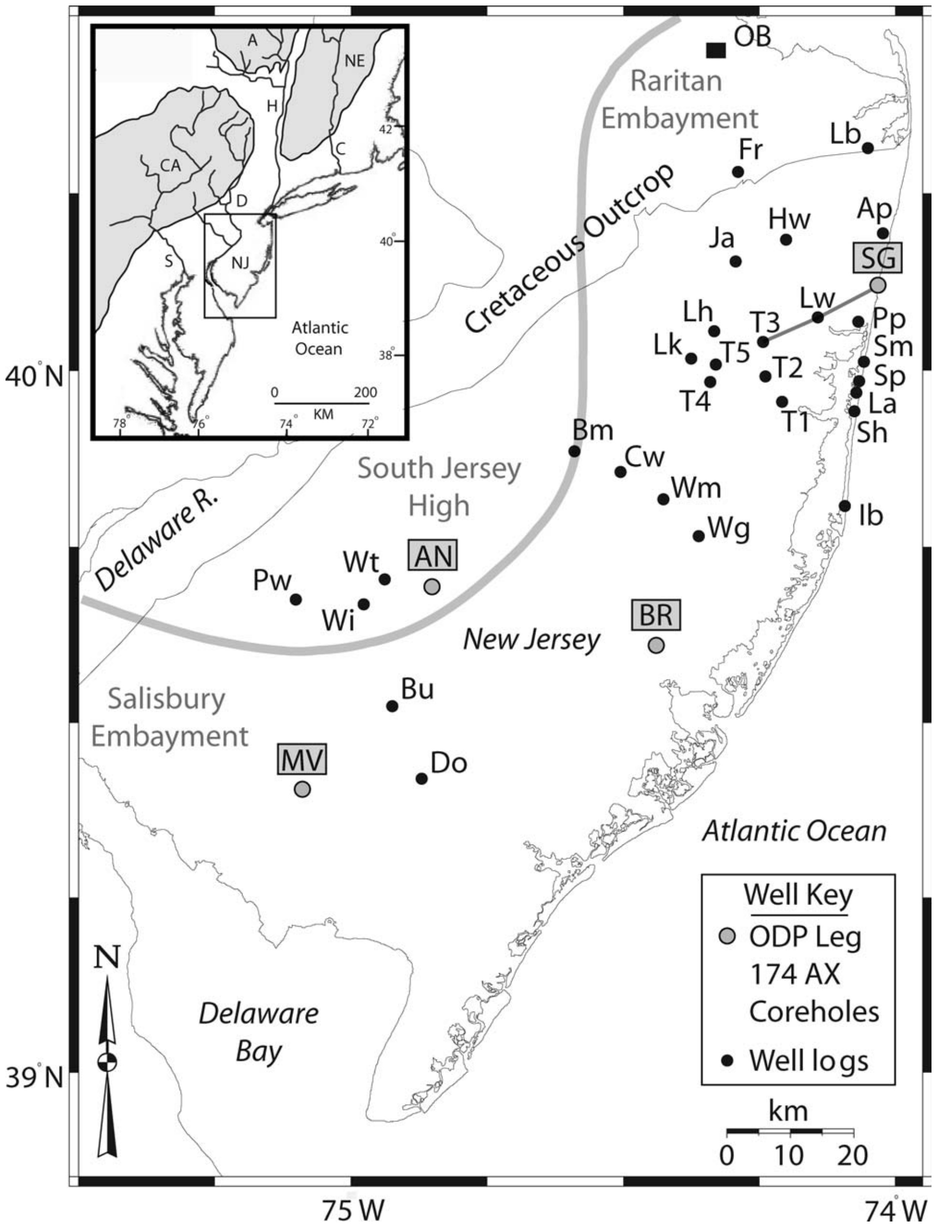

FIG. 1.—Location map shows the locations of ODP 174AX coreholes and additional geophysical logs used in this study. The location of basement structures is represented by the form line (in light gray) from Owens (1970). Coreholes (large gray circles, capital letters, boxed labels): AN, Ancora; BR, Bass River; MV, Millville; SG, Sea Girt. Geophysical logs (small circles, lowercase letters): Ap, Asbury Park; Bm, Browns Mills; Bu, Buena; Cw, Chatsworth; Do, Dorothy; Fr, Freehold; Hw, Howell; Ib, Island Beach; Ja, Jackson; La, Lavallette; Lb, Long Branch; Lh, Lakehurst; Lk, Lakehurst; Lw, Lakewood; Pp, Point Pleasant; Pw, Pittman West; Sh, Seaside Heights; Sm, South Mantoloking; Sp, Seaside Park; T1, T2, T3, T4, T5, Toms River area; Wg, Warren Grove; Wi, Williamstown; Wm, Woodmansie; Wt, Williamstown. OB represents an outcrop of the Magothy II sequence in Old Bridge, NJ. The inset map (after Poag and Sevon 1989) presents a regional view with modern river courses and source terrains (filled with gray): H, Hudson River; D, Delaware River; S, Susquehanna River; C, Connecticut River; CA, Central Appalachian Highlands; A, Adirondack Highlands; NE, New England Highlands; and NJ, New Jersey. The solid line represents a transect from Toms River to Sea Girt.

the response of deltaic facies, highstand sands, and eustatically forced depositional sequences to post-rift thermoflexural subsidence and higher-frequency autogenic fluvial axis switching and sediment supply variations; and (3) evaluate the relative influences of wave, tidal, and fluvial processes on deltaic sedimentation during the Late Cretaceous.

GEOLOGIC BACKGROUND

Upper Cretaceous sequences of the New Jersey Coastal Plain (Fig. 1) typically represent transgressive–regressive coarsening-upward successions from marine shelf to shallow marine, fluvio-deltaic, and nonmarine environments (Owens and Gohn 1985; Sugarman et al. 1995). Late Cretaceous (e.g., Olsson 1991; Miller et al. 2004) and Cenozoic (e.g., Miller et al. 1991; Miller et al. 1998a) strata record numerous unconformities (interpreted as sequence boundaries) caused by eustatic falls since the initial deposition of marine units during the Cenomanian (ca. 100 Ma) (Fig. 2) (Olsson et al. 1988; Miller et al. 2004). Coastal-plain sediments were deposited atop a series of basement structures: (1) the Salisbury Embayment, a large basin centered near Salisbury, Maryland; (2) the Raritan Embayment, located at the modern confluence of the Raritan and Hudson Rivers in Raritan Bay; and (3) the South Jersey High, a minor arch that divides the two embayments (Fig. 1) (Owens and Gohn 1985). The coastal plain was eroded during the global sea-level lowstands of the Plio-Pleistocene (Stanford et al. 2001), resulting in the exposure of Upper Cretaceous and Cenozoic strata.

The recognition of sequences as fundamental building blocks of the stratigraphic record greatly improved the understanding of New Jersey Coastal Plain stratigraphy and its controlling mechanisms. Owens and Sohl (1969) and Owens and Gohn (1985) identified transgressive–regressive coarsening-upwards cycles on the basis of recurrent glauconitic beds, physical unconformities, and biostratigraphic hiatuses and tied their cyclicity to regional tectonic processes. Olsson et al. (1988; Olsson 1991) identified and dated eight Upper Cretaceous New Jersey sequences and linked their origin to multiple Late Cretaceous marine transgressions

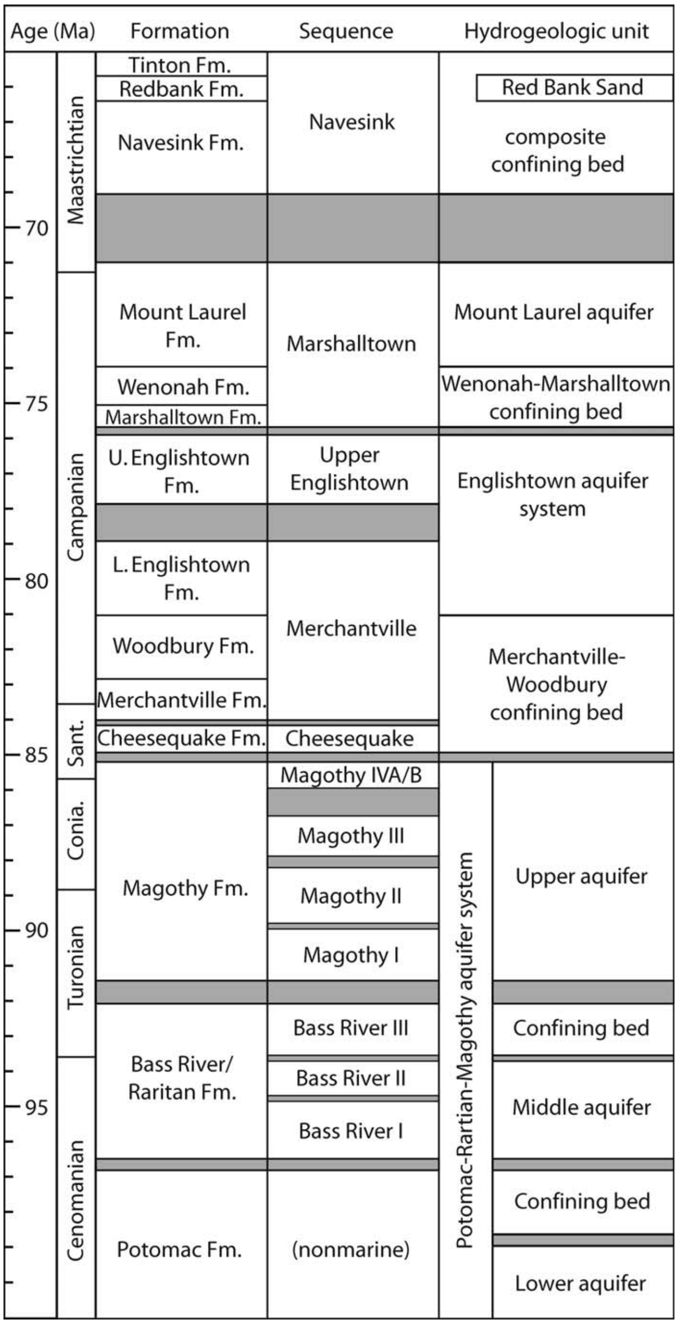

FIG. 2.—Generalized lithostratigraphy (after Owens and Gohn 1985), sequence stratigraphy (after Miller et al. 2004), and hydrostratigraphy (after Zapecza 1989) of the Upper Cretaceous New Jersey Coastal Plain.

(Albian through Maastrichtian), consistent with the eustatic control of Haq et al. (1987). Sugarman et al. (1995) integrated Sr-isotope stratigraphy with a biostratigraphic and lithostratigraphic framework to improve stratigraphic control ($< \sim 1$ my) of coastal plain sequences.

The New Jersey Coastal Plain Drilling Project drilled eleven onshore coreholes as part of the Ocean Drilling Program (ODP) New Jersey Sea Level/Mid-Atlantic Transect (NJ-MAT). ODP Legs 150 and 150X targeted Cenozoic sequences onshore at three sites and on the continental slope and established a link between Oligocene–middle Miocene sequence boundaries and glacioeustatic fall (Miller et al. 1996; Miller et al. 1991; Miller et al. 1998a). ODP Leg 174AX continued onshore drilling at eight sites, four of which targeted Upper Cretaceous strata (Bass River, Ancora, Millville, and Sea Girt) (Fig. 1). Results from these four coreholes identified and dated at least 11 (and as many as 18) sequences (Fig. 2) (Miller et al. 2003; Miller et al. 2004). Sequence boundaries in core are identified by: (1) a sharp unconformable contact; (2) lag gravels; (3) rip-up clasts; (4) extensive burrowing and bioturbation; (5) overstepping of facies successions; and (6) biostratigraphic and geochronologic hiatuses determined from benthic foraminiferal assemblages and Sr-isotope data (Miller et al. 2004; Sugarman et al. 1995). Although each sequence boundary is unique, together these methods can be used to indicate significant periods of nondeposition and erosion (Olsson et al. 1988; Sugarman et al. 1995). Water-depth variations within the sequences were established from lithofacies and biofacies analyses.

The studies by Miller et al. (2005) established the New Jersey margin as an excellent location for extracting estimates of global sea level for the Late Cretaceous and Cenozoic due to the well-preserved record of marine sediments and simple thermoflexural subsidence. One-dimensional backstripping (an inverse modeling technique that accounts for compaction, loading, subsidence, and paleodepth to determine accommodation rates and eustasy) indicates that large (> 20 m), rapid (< 1 Myr), and possibly glacioeustatic sea-level changes occurred during the Late Cretaceous (Miller et al. 2003; Miller et al. 2004; Miller et al. 2005; Van Sickel et al. 2004; Sugarman et al. 2005).

Although Miller et al. (2004) identified 11–16 Late Cretaceous New Jersey sequences and genetically linked them to eustasy, the understanding of subsurface sequence distribution is inherently limited due to the small number (four) of onshore coreholes that penetrate Upper Cretaceous strata and the large distances (~ 65 km) that separate coreholes. Without a detailed understanding of sequence expression across the coastal plain, the respective influences of global sea-level change, tectonic subsidence, and sediment supply remain clouded. In this study, geophysical logs bridge the gaps between coreholes, lending a sub-regional perspective to the distribution of coastal plain sequences, the paleogeography of deltaic facies systems, and a chronology of depocenter migration.

METHODOLOGY

Cores and Correlation of Geophysical Logs

This study uses gamma and electric logs from ODP Leg 174AX (Miller et al. 1998a; Miller et al. 1999; Miller et al. 2006; Sugarman et al. 2005) to establish a characteristic geophysical log signature for the sequences and lithofacies identified by Miller et al. (2004) and Lanci et al. (2002). This signature was used to identify sequences from logs lacking core control, allowing high-resolution (better than meter scale) mapping across the coastal plain, identification of sedimentary facies, and generation of a paleogeographic framework.

Sequence depocenters are identified by the thickest preserved intervals of a given sequence on the coastal plain. Although erosion occurs during base-level lowerings (forming the unconformities critical to this study), geographic variations in erosion do not appear to control the observed differences in thickness. This is established from: (1) similar age estimates of the sediments immediately above and below unconformities observed in core; (2) comparable thickness ratios of systems tracts within sequences; and (3) the preservation of well-developed upper highstand systems tracts (uHST) in most sequences, indicating that erosion was not severe enough to remove entire systems tracts.

Geophysical logs have been used to interpret paleoenvironments and correlate depositional facies since Serra and Sulpice (1975) used spontaneous potential (SP) and resistivity logs to unravel the depositional history of strata in the Gulf of Mexico. Gamma logs, a measure of naturally occurring radiation in sediment, have become a useful tool for log-based facies interpretation, particularly in siliciclastic fluvio-deltaic

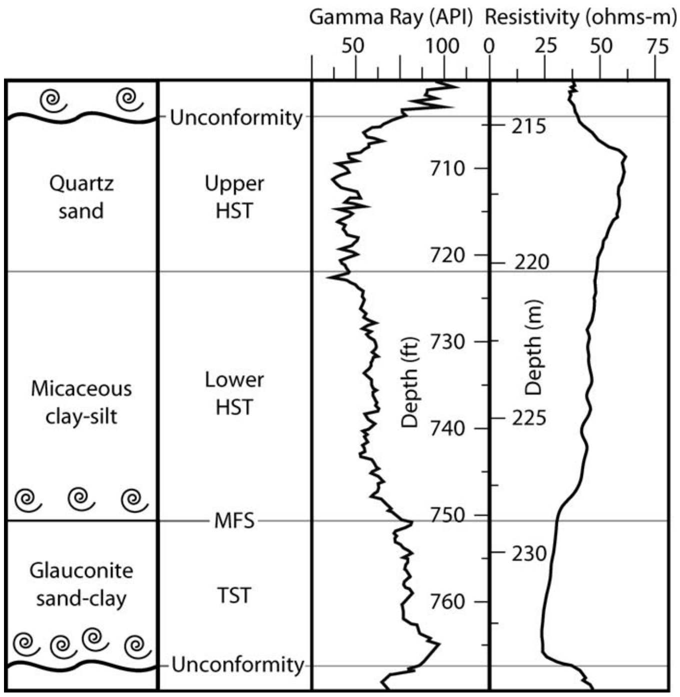

FIG. 3.— Anatomy and well-log signature of a typical New Jersey Upper Cretaceous sequence showing the primary lithologic components, their relationship to sequence stratigraphic units, and their gamma-ray and resistivity log characteristics (after Miller et al. 2004).

environments with good lithologic control from core, cuttings, or chip samples. Because fine-grained sediments, clays, glauconite sands, and phosphorites retain high levels of radiogenic elements, gamma logs are considered a good indicator of lithology (Rider 2002).

Within a transgressive–regressive sequence, gamma logs typically exhibit: (1) a sharp positive deflection across a basal unconformity; (2) high values (e.g., ~ 100–150 API units) in middle-neritic glauconite shelf sands and clays representing the transgressive systems tract (TST) of Posamentier and Vail (1988); (3) intermediate values (e.g., ~ 50–100 API units) in prodelta silty clays representing the lower highstand systems tract (lHST); (4) relatively low values (e.g., ~ 10–40 API units) in medium to coarse sands of the upper highstand systems tract (uHST); and (5) a rapid deflection to high values representing a sequence boundary and return to fine-grained glauconitic units (Fig. 3) (e.g., Lanci et al. 2002). Because the coarse upper-delta-plain or nearshore sands are also the primary groundwater aquifers of the coastal plain, resistivity logs (the measure of pore fluid resistance to an electrical current) generally exhibit high values (e.g., ~ 50–150 ohms-m) in coarser-grained intervals and significantly lower values (e.g., ~ 10–50 ohms-m) in finer-grained "confining" intervals (e.g., transgressive clays) (Fig. 3) (Keys and MacCary 1971). Although most sequences reflect the above patterns, varying sedimentation rates and levels of lowstand erosion can alter the expression of a sequence across the coastal plain (e.g., thin or absent highstand sands). Care must be taken to avoid oversimplified and incorrect interpretations (Rider 1990).

Gamma and electric logs have characteristic geometries that are useful for facies interpretation (Rider 2002). Gradual negative deflections capped by a sharp return to high gamma values, also referred to as "funnel" geometry, characterize a variety of sedimentary facies. These can consist of regressive shelf to delta front, prograding estuary, crevasse splay, and shoreface facies (Fig. 4) (Finley and Tyler 1986; Rider 2002). Conversely, a sharp negative shift overlain by a gradual positive deflection (a "bell" shape) can represent transgressive shelf, fining-upward fluvial channel (e.g., point bar), distributary channel, and wave-dominated delta front facies (Fig. 4) (Finley and Tyler 1986; Rider 2002). Trough-shaped low gamma values sharply bracketed by high gamma values can represent deltaic distributary and channel facies (Fig. 4) (Rider 2002). A "serrated" gamma signature can characterize swamp, marsh, lake, and levee facies (Finley and Tyler 1986; Rider 2002). Upper-delta-plain environments exhibit a variety of the log patterns discussed above, including floodplain paleosols that show high-amplitude, sharp, positive spikes in the middle of coarser-grained intervals (e.g., channel facies) (Fig. 4). By themselves, gamma and electric log interpretations are thus non-unique. In this study, we calibrate downhole logs and continuous cores (including lithologic and paleontologic control) to provide accurate paleoenvironmental interpretations that can be extended beyond core control to wells across the coastal plain.

Mapping

Detailed subsurface maps and cross sections of 13 Late Cretaceous sequences were generated using the sequence stratigraphic model of Miller et al. (1998a, 2004). In addition, 11 paleogeographic maps and accompanying sand thickness maps were created for sequences that exhibit shallow to nonmarine facies. Twenty-eight geophysical logs obtained from the New Jersey Geological Survey and industry sources were used to compliment the existing four ODP Leg 174AX coreholes at Bass River, Ancora, Millville, and Sea Girt (Fig. 1). Wells were selected for inclusion into the database on the basis of geographic location (e.g., satisfying areas of poor coverage), depth (substantial penetration through Upper Cretaceous section), and adequate quality.

Selected wells were required to include a gamma log, although it was preferred that they also include additional electric (primarily resistivity) logs. The combination of gamma and resistivity logs can offset the difficulty of correlating subtle lithologic changes in very fine-grained or glauconitic intervals (Fig. 5). If a sandy unit is not recorded on the gamma ray log, high resistivity readings might indicate its presence and prevent incorrect interpretation and correlation. Spontaneous potential and sonic logs provided an additional data source when gamma log correlation was unclear. Although this study relied heavily on geophysical log data as a method for correlation, physical and biostratigraphic data from ODP cores were used to constrain log signatures and account for sub-regional facies changes.

Downhole logs allow mapping of sandbodies within sequences. Sand isopach maps (depicted by 10 m contours on the paleogeographic maps) measured the total HST sand thickness per sequence. "Sands" were defined as intervals with gamma measurements lower than 75 API, although inconsistencies in older logs (acquired prior to the standardization of gamma tools) necessitated calibration to known measurements from the nearest corehole. In heterogeneous lithologies (e.g., lagoon, crevasse splay, delta plain), fine-grained intervals were subtracted from the total unit thickness to yield net sand thickness.

Low sedimentation rates, deep-water marine facies (less sensitive to base-level variations), unfossiliferous zones, and poor core recovery can complicate the identification and correlation of sequences. For this reason, coupled with the inherent limitations of resolution in detailed well log correlation, this analysis of Upper Cretaceous sequence distribution focused on mapping the most significant and pronounced sequences of the New Jersey Coastal Plain. As a result, the subdivisions of the Merchantville (I, II, III) and Navesink (I, II) sequences proposed by Miller et al. (2004) have been omitted because of their thin (< 10 ft, 3 m) expression in outer-shelf facies (e.g., contained entirely within intervals rich in glauconite and clay).

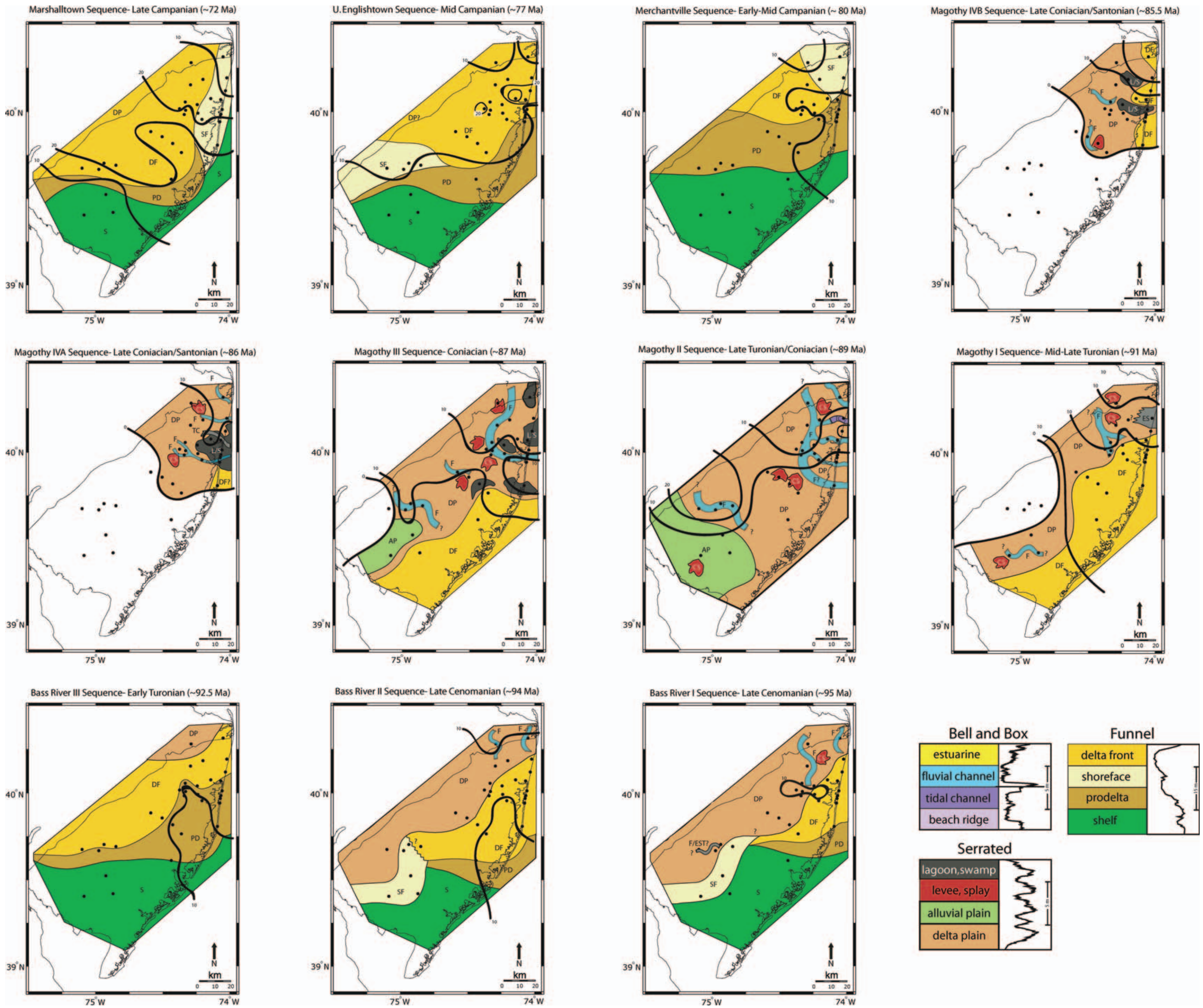

FIG. 4.—Paleogeographic maps showing the depositional evolution of Late Cretaceous deltaic facies. The thickness of highstand sands is represented by 10 m contour intervals. Facies are represented by both color and symbol where appropriate: AP, alluvial plain; CS, crevasse splay; DF, delta front; DP, delta plain and paleosols; ES, estuary; F, fluvial; L/S, lagoon and swamp; PD, prodelta; S, marine shelf; SF, shoreface; and TC, tidal channel to tidal delta. "Funnel, bell and box, and serrated" refer to the characteristic gamma log signatures of deltaic facies (listed to the left of each log) that are used to correlate facies away from continuous coreholes (after Rider 2002). Approximate vertical scales are displayed to the right of each log signature.

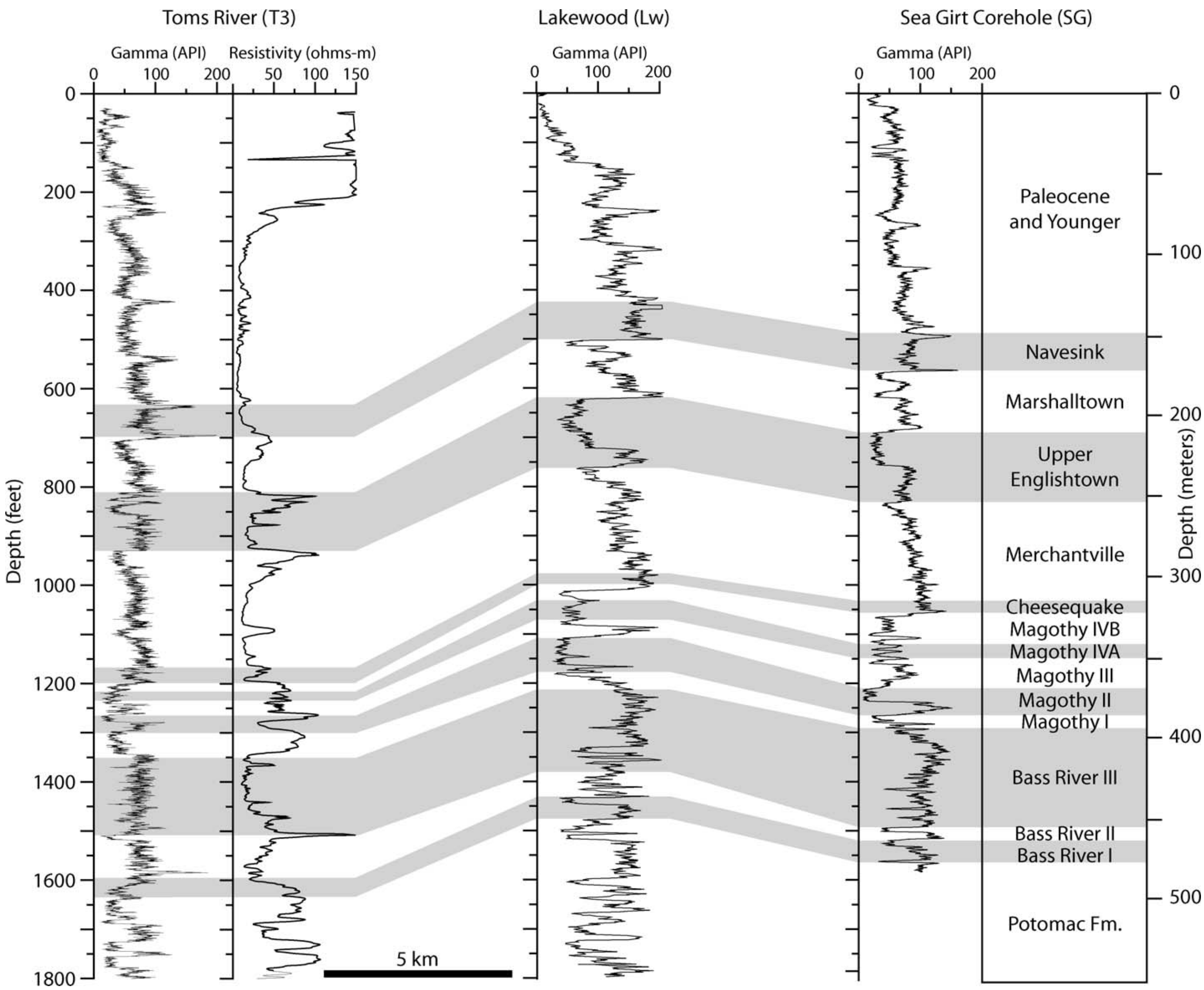

FIG. 5.—Strike cross section of Upper Cretaceous sequences of the northern New Jersey Coastal Plain showing typical well-log characteristics. Correlation between ODP 174AX corehole Sea Girt (SG) and geophysical logs Lakewood (Lw) and Toms River (T3) shows the advantage of using both resistivity and gamma-log data (e.g., T3), particularly in identifying thin HST sand units. Location of cross section shown on Fig. 1.

DATA AND RESULTS

Early Cenomanian–Early Turonian Sequences

The identification of characteristic geophysical log signatures for sequences and facies in New Jersey is unique to this study and enabled the mapping of sequences first identified by Miller et al. (2004) across the coastal plain. To emphasize the critical link between sequence components, lithofacies, and log character, previously published sequence descriptions are referenced for each sequence. Three sequences (Bass River I, II, III) were identified in the Bass River Formation (Sugarman et al. 2005). The Bass River I sequence is dated as early to mid-Cenomanian (Pollen Zone IV) and unconformably overlies the fluvial and terrestrial mottled clays and paleosols of the Barremian–lowermost Cenomanian Potomac Formation. The Bass River II sequence is mid-Cenomanian whereas Bass River III, the uppermost and thickest of the sequences, is upper Cenomanian–lower Turonian (Fig. 2) (Miller et al. 2004). Bass River sequences generally "shallow" upwards from: (1) neritic glauconite sand and clay (TST); (2) prodelta clay and silt (lHST); and (3) delta front to shoreface quartz sands (uHST) (Miller et al. 1998b; Miller et al. 1999; Miller et al. 2006; Sugarman et al. 2005).

Each sequence is characterized by a coarsening upwards "funnel" gamma log signature, although local variations can result in a "serrated" or "box" character (Fig. 5). The lack of thick upper HST clean coarse quartz sands often results in relatively high gamma values, highlighting the importance of resistivity logs and their peak values in these thin, sandy water-bearing intervals to identify highstand deposits.

Isopach maps of the Bass River sequences reveal pervasive, downdip (seaward) thickening toward the central and southern coastal plain (Fig. 6). Bass River I is 10–15 m thick across most of the coastal plain, although: (1) several coastal sections exceed 15 m; (2) a comparatively thick "finger" (+ 15 m) extends updip from the coast towards central New Jersey; and (3) the thickest interval (~ 21 m) is located in the southern coastal plain (Fig. 6). Highstand delta front sands are thickest in the central coastal plain (10.9 m), but grade to shoreface sands ~ 1–2 m thick to the south (Fig. 4). The Bass River II sequence is thick across the coastal plain (~ 24 m) but thins to the west at Ancora (~ 8–10 m) and

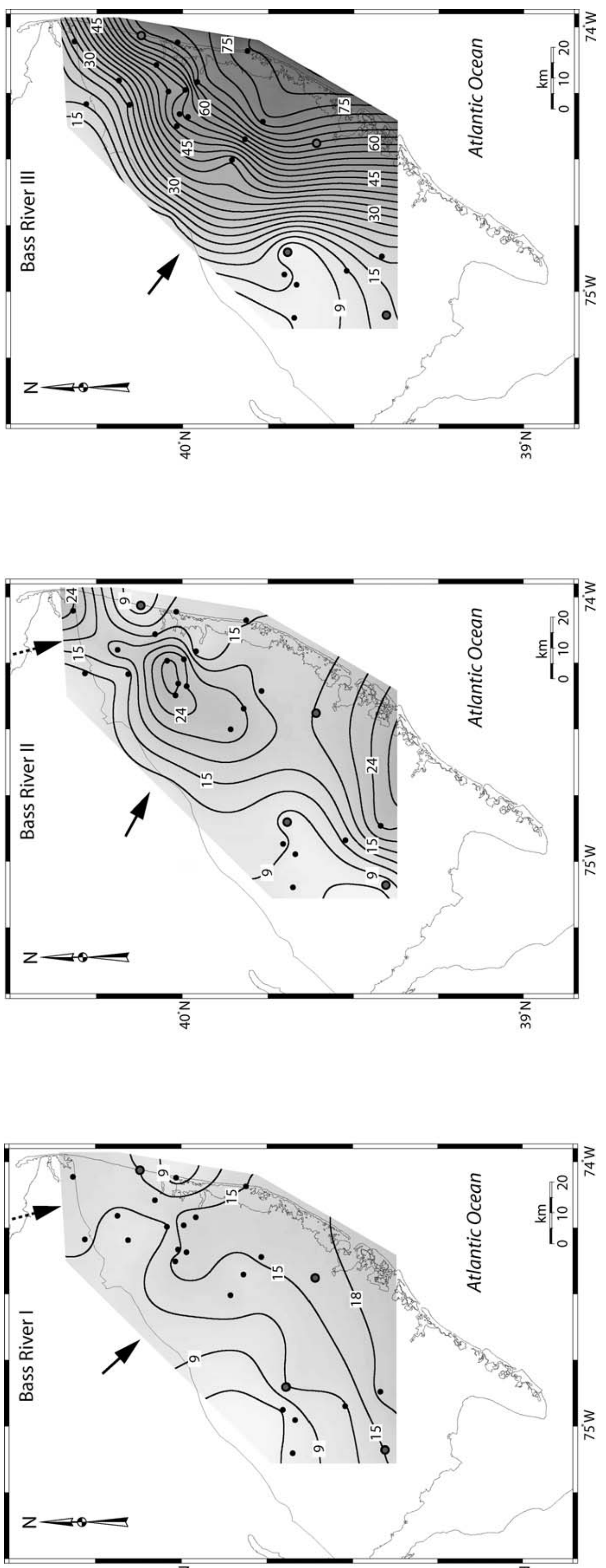

FIG. 6.—Isopach maps of the Cenomanian Bass River sequences (I, II, III) from core and geophysical log data. Arrows represent the inferred locations of sediment sources, with solid arrows representing a primary source and dashed arrows indicating a weaker secondary source. Contour interval is 3 meters. Large gray circles represent coreholes; small black circles represent well locations.

to the south at Millville (9.7 m). Thick values (24–27 m) in the central coastal plain could represent a localized depocenter during the mid-Cenomanian (Fig. 6). Highstand sands range from 10 to 11.5 m across the central and northern coastal plain and vary from delta-front to fluvial in origin. Thin (1–3.5 m) delta-plain sands dominate the southern coastal plain. The Bass River III is the thickest (and most variable) Cenomanian sequence inasmuch as it: (1) thickens to 77.7 m at Island Beach; (2) thins to the west (6–8 m); and (3) thins to the south at Millville (14.1 m). Highstand sands thicken to 10 m at Bass River, but are otherwise thin (1–5 m) and fine-grained across much of the coastal plain.

Mid-Turonian–Coniacian Sequences

Five Turonian–Coniacian sequences (Magothy I, II, III, IVA, and IVB) were identified in the coarse-grained Magothy Formation, one of New Jersey's primary aquifers and the upper unit of the PRM aquifer system (Fig. 2) (Zapecza 1989). Correlation using pollen zonation after Christopher (1982) reveals a discontinuous and patchy distribution for Magothy sequences. The lowermost Magothy I is dated as pollen Zone IV (mid-Turonian) at Bass River, Millville, and Sea Girt (Miller et al. 1999; Miller et al. 2006; Sugarman et al. 2005). The Magothy II, the oldest sequence recovered at Ancora, is assigned to pollen Zone V (late Turonian), indicating that the Magothy I was cut out (Miller et al. 1999). The Magothy III sequence is assigned to pollen zone V and was recovered at all four coreholes (Miller et al. 1998; Miller et al. 1999; Miller et al. 2006; Sugarman et al. 2005). Two Coniacian (Pollen Zone VII) Magothy sequences (IVA and IVB) were penetrated only at the Sea Girt corehole and are restricted to the northern coastal plain (Miller et al. 2006).

The Sea Girt corehole provides an expanded Turonian–Coniacian section that allows the first core and log samples of the IVA and B sequences near local outcrops of the Magothy Formation (Miller et al. 2006). Sugarman et al. (2006) informally linked the recovered sequences (pending further analyses) to outcrop-defined members of the Magothy Formation (Fig. 7) on the basis of lithology: the (1) Sayreville Sand Member (Magothy I); (2) South Amboy Fire Clay and Old Bridge Sand members (Magothy II); (3) Amboy Stoneware Clay Member (Magothy III); (4) the Morgan Beds Member (Magothy IV A); and (5) the Cliffwood Beds Member (Magothy IVB).

Although Magothy sequences consist of a diverse system of deltaic facies (Fig. 7), they maintain fairly consistent well-log signatures across the coastal plain (Fig. 5). The delta-front and fluvial sands of the Magothy I and III (Miller et al. 1998b; Miller et al. 1999) are easily distinguishable by significant gamma-ray troughs (and high resistivity values) separated by the high gamma peaks of the Magothy II clays and paleosols (Fig. 5). In the northern coastal plain, the delta-front sands of Magothy IVA form four distinct gamma spikes that resemble a "serrated" log pattern (Fig. 7). The overlying delta-front and bay to lagoon sands of the Magothy IVB sequence exhibit a similar low amplitude "choppy" gamma interval capped by a sharp deflection to high values of the overlying Cheesequake Formation. The Magothy IV sequences maintain these signatures until they gradually pinch out in the central coastal plain (Fig. 4).

Thickness trends of Magothy sequences fall into two distinct groups: (1) the lower (Magothy I, II, III) sequences that are present throughout the coastal plain (Fig. 8); and (2) the upper (IVA, IVB) sequences that are limited to the north (Fig. 8). The Magothy I sequence is widespread but discontinuous, grading from thick intervals in the central coastal plain (18.6 m) to the point of no recovery at Ancora. Northern sections show a patchy distribution from 7.6 to 15.2 m and the sequence thins to 3.4 m in the south (Fig. 8). Thick highstand sands are concentrated in the central (12–18 m) and northern (14.6 m) coastal plain and range from delta front to crevasse splay (Fig. 4). Sands become significantly thinner (> 1 m) towards Millville. The Magothy II sequence reveals two primary

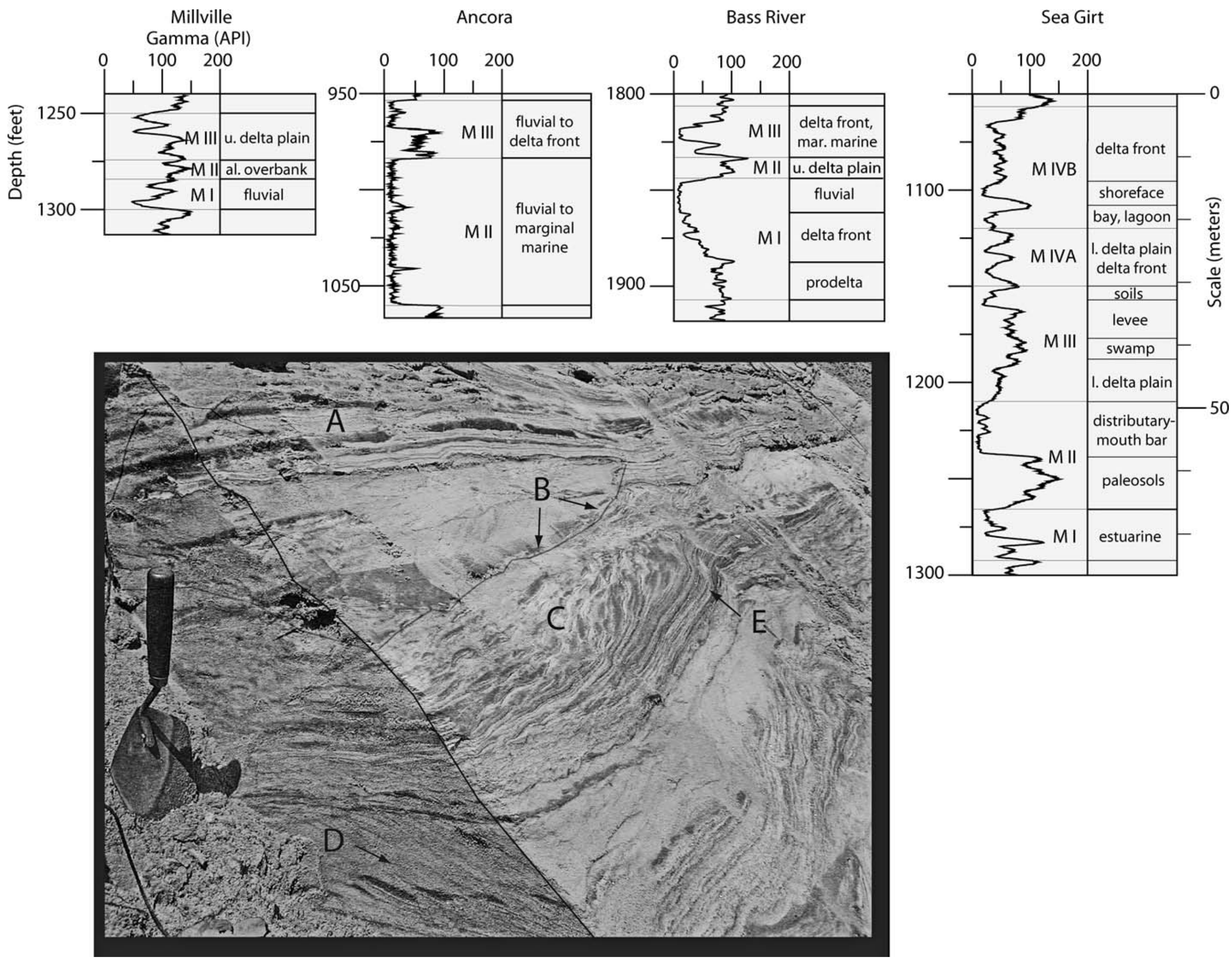

FIG. 7.—Subdivisions of the Magothy Formation showing sequences, facies, and log characteristics from the four ODP 174AX coreholes. Magothy sequences thicken to the north and contain diverse shallow marine to delta-plain facies. Abbreviation of facies: al, alluvial; u, upper; l, lower; mar, marginal. Outcrop photograph of the Old Bridge Member of the Magothy Formation (Magothy II sequence), Old Bridge, New Jersey. Note the change in orientation of the outcrop face represented by the thin sub-vertical black line. The letters indicate: **A)** an irregular-based interval of gray to white clean clays with fine-grained quartz sand; **B)** an irregular-based (30–40 cm) channel incised into clayey and sandy substrates; **C)** flaser and wavy beds of black to dark gray micaceous, organic clay draped over small (1–2 cm) ripples and planar cross-beds of fine- to medium-grained quartz sand; **D)** planar cross stratification; and **E)** interlaminated (3–5 mm) black micaceous clay with yellow to gray fine quartz sands. Handle of trowel is 15 cm for scale. This figure is in color in the digital version of the journal.

depocenters: (1) a thick section in the northeast reaches 18.3–21.3 m; and (2) a thick western interval measures + 26 m at Ancora (Fig. 8). Thick highstand fluvial sands (16–25 m) at Ancora are consistent with these observations. The Magothy III sequence thickens eastward from thin (Ancora: 7.9 m) or absent to thick intervals at Sea Girt (19.7 m) and Dorothy (17.2 m). Intermediate sections (9.1–13.7 m) characterize much of the central coastal plain. The thickest highstand sands (9–14.6 m) are found around Ancora and throughout the central coastal plain.

The upper Magothy (IVA, B) sequences are restricted to the northern coastal plain and were only recovered at the Sea Girt corehole, though logs allow correlation in the north (Fig. 8). The Magothy IVA thickens to 17.5 m in the northeast and thins consistently to the south before pinching out in the central coastal plain (Fig. 8). The Magothy IVB thickens to twin northern ‘‘bulls-eye’’ depocenters at Sea Girt (17.8 m) and Freehold (16.2 m) and is similarly absent from the southern coastal plain (Fig. 8). Patterns of highstand sand thickness are consistent with overall sequence thickness trends for the Magothy IVA and IVB and represent an array of fluvial channel, delta-front, and lagoonal sands (Fig. 4).

Santonian Sequence

A comparatively thin (8–26 ft, 2.4–7.9 m) lower to middle Santonian Cheesequake sequence is identified in cores (Miller et al. 1998b; Miller et al. 1999) and correlated to the glauconitic clayey silt of the Cheesequake Formation of outcrop (Fig. 2) (Owens et al. 1998). The Cheesequake Formation and sequence is dominated by inner-shelf to middle-shelf facies and bracketed by distinct unconformities with the Magothy and Merchantville formations (Fig. 9) (Miller et al. 2004).

The Cheesequake sequence is correlated across the coastal plain on the basis of its geophysical log signature and position between the glauconitic clays of the Merchantville Formation (high gamma values above) and the

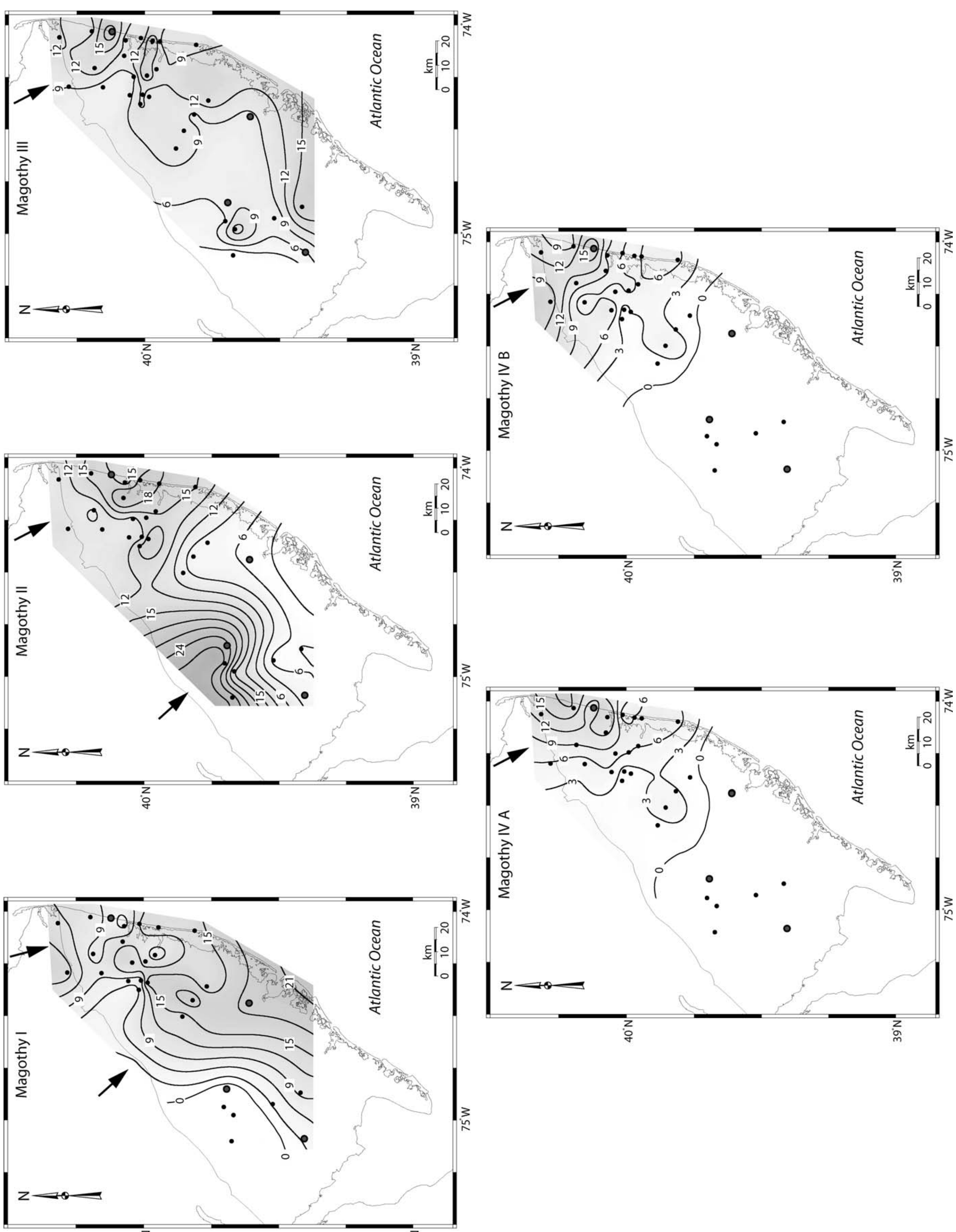

FIG. 8.—Isopach maps of the Turonian–Coniacian Magothy sequences (I, II, III, IVA, IVB) from core and geophysical log data. Solid arrows represent a primary sediment source and dashed arrows indicate a weaker secondary source. Contour interval is 3 meters. Large gray circles represent coreholes; small black circles represent well locations.

coarse quartz sands of the Magothy Formation (low gamma, high resistivity below). Dual gamma spikes often mark the gradation from basal glauconitic clays and sands (TST) to fine quartz sands (HST), although a series of smaller peaks within the sequence represent the interplay of clayey glauconite beds and coarser lithology (Fig. 5).

The Cheesequake sequence exhibits gentle, downdip thickening from 3–4.2 m to its maximum thickness of 7.9 m at Bass River. The sequence thins to the south (2.4–3.4 m), possibly representing the influence of the South Jersey High. No significant onshore depocenters or significant quantities of shallow marine highstand sands are apparent during the Santonian, a reflection of deposition on a sediment-starved shelf with weak northern and southern sources. For this reason, paleogeographic and deltaic facies maps were not created for the Cheesequake sequence.

Uppermost Santonian–Campanian Sequences

Three predominantly Campanian sequences were identified from core: (1) the uppermost Santonian to mid-Campanian Merchantville sequence; (2) the mid-Campanian upper Englishtown sequence; and (3) the upper Campanian Marshalltown sequence (Fig. 2) (Miller et al. 2004). Unlike Bass River and Magothy sequences, some Campanian sequences encompass multiple Upper Cretaceous lithostratigraphic units. The Merchantville sequence consists of the Merchantville, Woodbury, and lower Englishtown formations (Miller et al. 2004). The upper Englishtown sequence corresponds to the upper part (informal) of the Englishtown Formation (Owens et al. 1998), and the Marshalltown sequence consists of the Marshalltown, Wenonah, and Mount Laurel formations (Fig. 2) (Miller et al. 2004).

The Merchantville sequence exhibits classic funnel geometry on the gamma log (Fig. 5) and a "coarsening upwards" gradational succession of glauconite clay and sand, micaceous clay, and fine quartz sand (Miller et al. 1998b; Miller et al. 1999). The HST of the sequence (lower Englishtown Formation) is a moderate coastal-plain aquifer (Zapecza 1989) and exhibits high resistivity values (Fig. 2). The Merchantville sequence thickens downdip and northeast from thin western sections (46.5 m) to 89.3 m at Island Beach, the thickest upper Cretaceous sequence observed on the coastal plain. Thick intervals (60.9–73.1 m) are visible across the central and northern coastal plain, but the sequence thins towards the South Jersey High (Fig. 9). Highstand sands are thickest along the central coast and inland (10–14 m) and consist of interbedded delta-front and prodelta deposits. Shoreface sands in the north range from 6.7 to 8 m, whereas the southern coastal plain exhibits thin (3–5 m) shelf sands (Fig. 4).

The geophysical log signature of the upper Englishtown sequence, an important northern aquifer composed largely of quartz sand, varies significantly along dip. In the northern coastal plain, thick delta-front sections are easily distinguished by their "box-like" appearance in gamma logs and high resistivity values (Fig. 5). The sequence thins consistently to the south around Millville and becomes increasingly fine-grained, glauconitic, and assumes a gradational "funnel" gamma signature. Although resistivity values are "muted" in these fine-grained intervals, the upper sand of the sequence prevails across the coastal plain. The upper Englishtown sequence is thickest in the northeastern coastal plain around Sea Girt (45.7–51.8 m) and thins west of Toms River (33.5 m). Highstand sands are generally 20–29 m thick around Sea Girt and grade to $\sim$ 10 m around the central coastal plain (Fig. 4). Thinning of the sequence (8.3 m) and highstand sands (6 m) around Millville could indicate increasing distance from a strong deltaic northern source (Fig. 9).

The Marshalltown sequence shallows upward from glauconite clay and sand, to micaceous clay and clean quartz sands (Miller et al. 1998b; Miller et al. 1999). This results in "funnel" geometry for gamma and resistivity values. The largest gamma spike (+ 150 API units) of the New Jersey Coastal Plain marks the sequence boundary between the Mount Laurel HST sand (upper Marshalltown sequence) and the glauconitic Navesink sequence (Fig. 5). This reworked interval is glauconite-rich and includes concentrations of phosphate pebbles, rip-up clasts, and represents a lowstand systems tract (LST) lag deposit of the overlying Navesink sequence (Miller et al. 2004). The Marshalltown sequence thickens seaward and exhibits two primary depocenters (central and northern) divided by a thin interval (21.3–27.4 m) (Fig. 9). The northern depocenter thickens to 43.6–47.4 m along the northeastern coast, whereas the southern depocenter is thickest at Bass River (44.5 m) and gradually thins to 32.3 m at Ancora. Very thin intervals characterize the southern coastal plain (8.1 m at Millville), just 30 km south of Bass River (Fig. 9). Thick highstand delta-front sands (21–25.6 m) occur across the central coastal plain but become finer-grained and thin to $<$ 1 m shelf sands at Millville. Deposition during the Campanian indicates a north-northeast shift in the sedimentation of the New Jersey Coastal Plain similar to Turonian–Coniacian trends (Magothy sequences) (Fig. 8). Cenomanian–lower Turonian sequences exhibit depocenters in the central and southern coastal plain, whereas Turonian–Campanian sequences thicken towards the north.

Maastrichtian Sequence

The Maastrichtian to lowermost Danian Navesink sequence consists of fossiliferous glauconite clays and sands (Fig. 2) (Miller et al. 1998b; Miller et al. 1999). This interval is characterized by high gamma values and low resistivity values, although sandier intervals (e.g., Redbank, Tinton formations) may exhibit slight resistivity peaks (Fig. 5). The Cretaceous–Paleogene (K–P) impact event is preserved in most Navesink sections and is marked by spherule layers, Cretaceous chalk fragments, and a major gamma peak that is only exceeded by the uppermost layer of the Mount Laurel sand (Olsson et al. 1997; Miller et al. 1998a). These two peaks dominate the gamma log signature and allow easy identification of the Navesink Formation, although reworking and bioturbation of the K–P boundary can obscure the identification of the upper sequence boundary (Fig. 5).

Thickness variations of the Navesink sequence appear largely unrelated to dip. The northernmost wells of the study area represent the thickest intervals (26.2 m). An east–west-trending band of thin Navesink (9.1–10.1 m) in the south central coastal plain divides similarly thick (13.7–18.2 m) southern and central deposits (Fig. 9). This relatively thick southern section is not consistent with the majority of Cenomanian–Campanian sequences that thin south of the Bass River corehole. The lack of a clear depocenter paired with relative thickening to the south could indicate: (1) deep-water environments with low sediment input; and/or (2) decreased influence of basement structure on margin deposition. Paleogeographic and facies distribution maps were not created for the Navesink sequence due to the abundance of shelfal facies and absence of shallow marine highstand sands.

DISCUSSION

Deltaic Facies Models

The analysis of deltaic systems has long been defined by the tripartite system of Galloway (1975), who used the relative influence of tidal, wave, and fluvial processes to classify delta morphology. Subsequent studies have shown that deltas and facies arrangements evolve through a broad spectrum of stages as a function of changes in sedimentation rates, eustasy, and rates of accommodation. Bhattacharya and Giosan (2003) show that deltas can exhibit wave-, tide-, and river-dominated facies across different lobes (e.g., Danube delta), suggesting that different classifications of delta can exist under the similar conditions (e.g., microtidal) within a delta system. Furthermore, deltas can evolve from

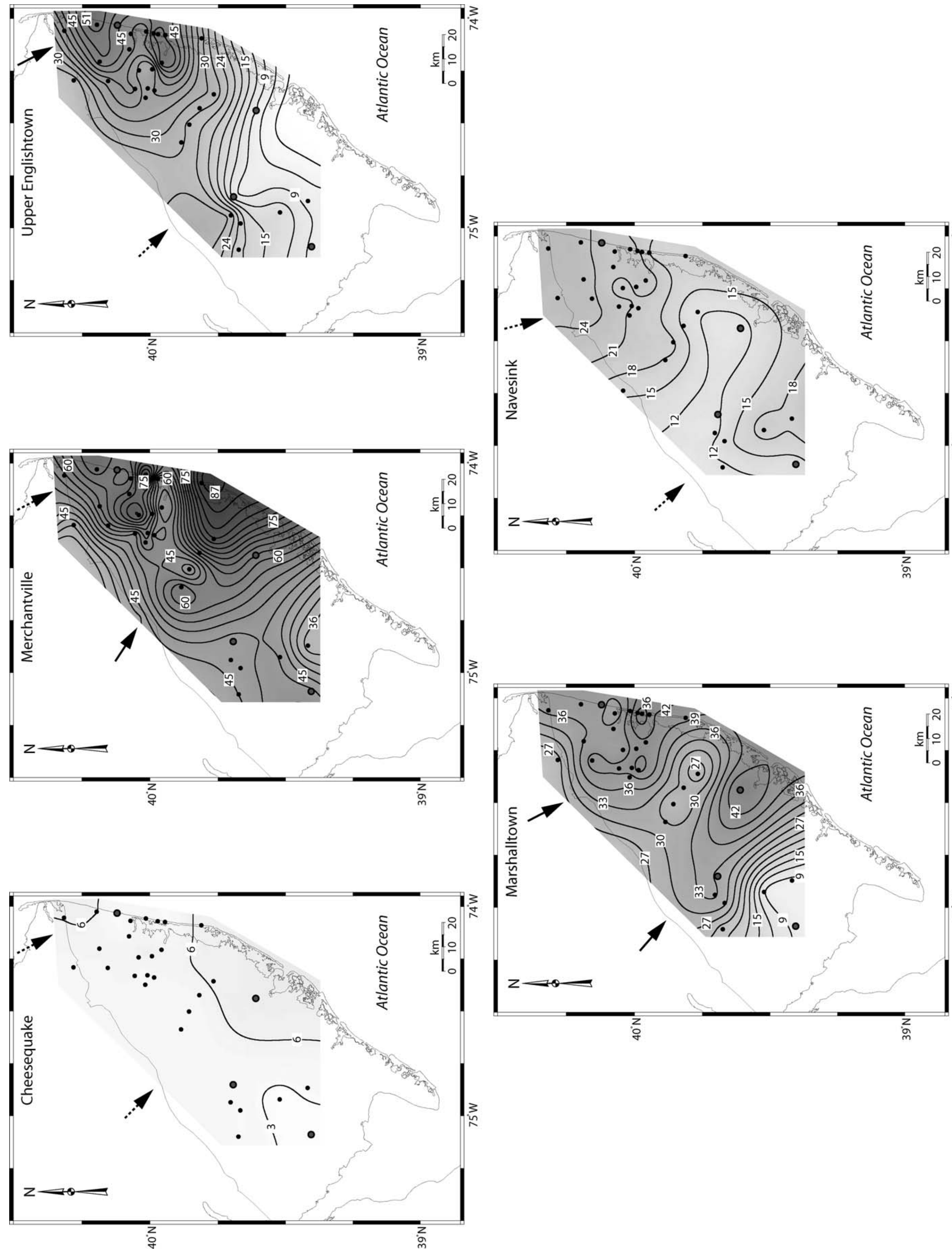

FIG. 9.—Isopach maps of the Santonian, Campanian, and Maastrichtian sequences (Cheesequake, Merchantville, upper Englishtown, Marshalltown, and Navesink) from core and geophysical log data. Solid arrows represent a primary sediment source and dashed arrows indicate a weaker secondary source. Contour interval is 3 meters. Large gray circles represent coreholes; small black circles represent well locations.

tide- to wave-dominated over relatively short time periods. The Mekong delta shifted from tide-dominated to tide- and wave-dominated over the last 4 kyr (Ta et al. 2002), while seasonal variations in wind and wave energy can also influence facies characteristics (Yang et al. 2005). It is therefore simplistic to assume that 35 Myr of Late Cretaceous sedimentation can be exclusively pinned to a three-end-member modern analog system defined by deposition during global sea-level highstands.

Despite these limitations of delta classification, wave, fluvial, and tidal processes distinctly influence facies deposition, sedimentary characteristics (on a variety of scales), and the distribution and geometry of subsurface units. Lithofacies analysis from continuous core, coupled with our use of geophysical logs to establish paleogeographic maps and the lateral relationships of deltaic facies, reveals the widely known variability of deltaic systems. However, we also document the relative stability of deltaic facies systems on the 10^6–10^7 yr scale (Fig. 10), with long periods of cyclically repeating systems tracts controlled by eustasy punctuated by facies shifts controlled by long-term sea level and shifting fluvial–deltaic sources (Fig. 4).

Early studies (Owens and Sohl 1969; Owens and Gohn 1985) recognized the deltaic origin of Upper Cretaceous New Jersey Coastal Plain strata, and subsequent lithofacies analyses by Miller et al. (2004) tied the observed shelf, prodelta, and shallow marine facies to a "mixed" tide- and wave-influenced modern Niger delta facies model (Allen 1970). Characteristics of these deltaic facies that we observe in core are: (1) thin middle-neritic to outer-neritic glauconite sands and clays (60–200 m paleodepths determined from benthic foraminiferal analysis); (2) common, thick prodelta micaceous clays and silts (20–60 m paleodepths); (3) generally thick delta-front, nearshore, and shoreface fine to coarse quartz sands (0–20 m paleodepths); (4) delta-plain sands, silts, and clays; (5) fine- to medium-grained fluvial, estuarine, and tidal-channel quartz sands; (6) back-barrier lagoon and swamp organic-rich clays and sands; (7) levee and crevasse-splay sands; (8) upper-delta-plain and lower-delta-plain interfluvial mudplain clays and paleosols (Miller et al. 1998b; Miller et al. 1999; Miller et al. 2003; Miller et al. 2004; Miller et al. 2006; Sugarman et al. 2005). The middle-neritic to outer-neritic facies compose the basal TST packages observed in the Bass River, Cheesequake, Merchantville, Marshalltown, and Navesink formations (Miller et al. 2004). Prodelta facies compose the lower HST found in the Bass River, Woodbury, upper Englishtown, and Marshalltown formations. Coarse-grained delta-front to shoreface facies represent the upper HST observed in the Bass River, lower Englishtown, upper Englishtown, and Mount Laurel formations (Miller et al. 2004). A majority of the marginal-marine to nonmarine facies are restricted to the Magothy Formation (Fig. 7), although they are occasionally observed in other sections (Miller et al. 2004).

Discerning the influence of wave, fluvial, and tidal processes on deltaic sediments in fully marine units can be difficult. Instead, careful examination of marginal to shallow marine facies proximal to the littoral zone can offer a snapshot of the processes that shape deltaic facies patterns. The paleogeographic distribution of facies can also be very useful in determining the relative influence, inasmuch as tidally influenced sandbodies tend to be shore perpendicular, while wave-dominated sandbodies are often arcuate, shore parallel, and show evidence of wave reworking and lateral transport by longshore currents (e.g., Van Andel 1967; Fisher and McGowan 1969; Allen 1970).

Paleogeographic maps reveal several examples of wave influence on sandbody geometry and deltaic facies patterns: (1) Thick (+ 20 m) delta-front sands of the Marshalltown sequence (Mount Laurel Formation) grade rapidly ($>$ 10–20 km) across the central coastal plain into similarly thick (21 m) shoreface sands to the northeast. These sands become progressively thinner farther north along the paleoshoreline, indicating increasing distance from the primary sediment input (Fig. 4). (2) A similar transition is visible in the upper Englishtown sequence, although shoreface sands are visible to the southwest of the main depocenter. Thick 10–29 m delta-front deposits of the central and northern coastal plain transition to thinner ($\sim$ 10 m) shoreface sands to the south at Ancora, although this lateral facies change occurs over $\sim$ 40 km (Fig. 4). (3) Thin ($<$ 10 m) delta-front and shoreface deposits of the Merchantville sequence (lower Englishtown Formation) are juxtaposed in the northern coastal plain around Sea Girt and Toms River. This transition occurs over a very short distance ($<$ 5–10 km), with shoreface sands becoming more abundant to the northeast (Fig. 4). (4) Both the Bass River I and II sequences exhibit delta-front sands in the central coastal plain that pass into thin shoreface sands in the southern coastal plain (Ancora and Millville area). Because this facies change occurs over the course of 40–50 km, discerning the nature of wave-reworking becomes difficult. (Fig. 4). While these shore-parallel transitions from thick delta-front deposits to shoreface sands could simply represent the contrast between deltaic and interdeltaic segments of a margin (particularly facies shifts that occur over long segments of a coastline), we believe that the rapid scale of these changes ($>$ 10–40 km) and comparable thicknesses of the delta-front and shoreface facies attest to wave reworking and consequent redistribution of sand by longshore drift.

While paleogeographic maps are useful in determining longshore variations in facies character and potential wave influence, discerning the orientation of tidally influenced sandbodies exceeds the spatial resolution of this study due to the geographic distribution of wells and coreholes. Determining the tidal influence of a sedimentary unit can be very difficult from core, necessitating further integration with outcrops. However, because most of the Late Cretaceous deltaic facies are either fully marine or nonmarine, few candidate sequences with nearshore to marginal marine facies are available for extracting tidal influence.

Observations from an outcrop of the Magothy II sequence in Old Bridge, NJ ($\sim$ 30 km to the northwest of the Sea Girt corehole) reveal sedimentary characteristics consistent with strong tidal influence on deposition. These include: (1) abundant flaser and wavy beds of black to dark gray micaceous, organic clay draped over small (1–2 cm) ripples and planar cross beds of fine- to medium-grained quartz sand; (2) interlaminated (3–5 mm) black micaceous clay with yellow to gray fine quartz sands; (3) numerous sets of large (50–100 cm thick) trough and planar cross beds interbedded with intervals of clean gray to white clays; (4) 1–2 cm diameter scours and irregular reactivation surfaces; (5) irregular-based (30–40 cm to 2 m) channels incised into clayey and sandy substrates; (6) rare bidirectional cross bedding with clay drapes; and (7) presence of rare 4–8 cm long *Skolithos* burrows, hinting at a marginal-marine environment of deposition (Fig. 7). These criteria support our interpretation of either a tidal channel or tidal delta for the Magothy II sequence at this locality.

Although the Old Bridge outcrop of the Magothy II offers only a brief snapshot of 35 Myr of Late Cretaceous deltaic sedimentation, facies identified throughout the Magothy sequences in core and from logs (e.g., tidal channel) support the identification of tidal influence. The presence of tidal channels, estuarine deposits, and extensive lagoons and swamps, coupled with broad sandbody trends derived from paleogeographic maps of the Turonian–Coniacian, is consistent with a mixed wave- and tide-dominated delta. The absence of marginal-marine facies throughout the Campanian, Santonian, and Cenomanian makes the identification of tidal influence difficult, but does not preclude it.

Paleogeographic Evolution of Late Cretaceous Deltas, New Jersey Margin

The paleogeographic reconstruction of Late Cretaceous deposition in the New Jersey Coastal Plain was generated using sequence boundaries as geochronologic markers. Facies analysis of the deposits directly underlying these sequence boundaries provided the geographic distribution of deltaic facies during the regressive highstand systems tract. A series of

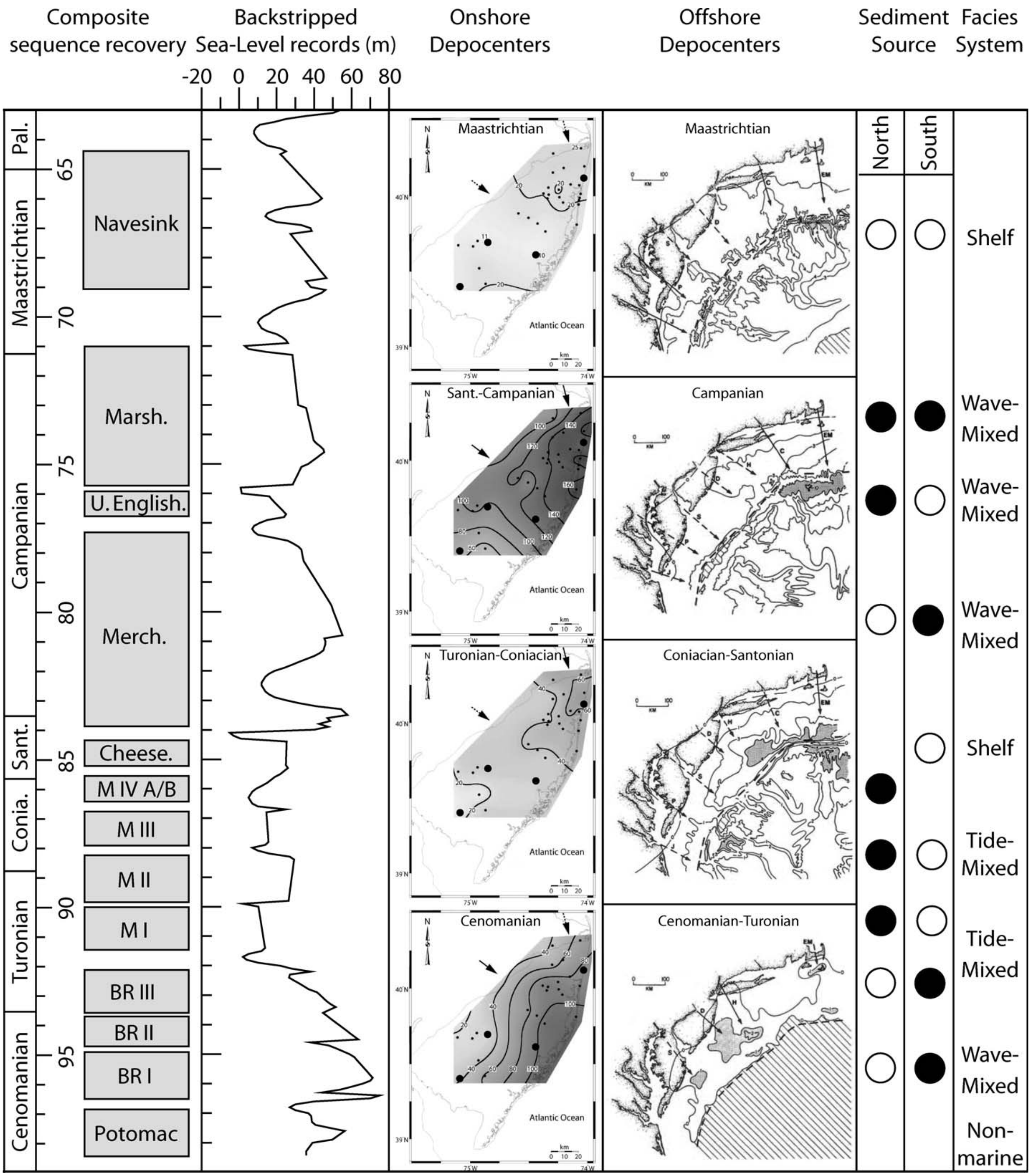

FIG. 10.—Chart shows: (1) Composite sequence recovery; (2) backstripped sea-level estimates with inferred lowstards (from Miller et al. 2005); (3) onshore depocenter isopach maps (contour interval 20 meters); (4) offshore depocenter isopach maps from Poag and Sevon (1989) (contour interval 100 meters); (5) inferred sediment source (black circle indicates primary role, open circle indicates secondary); and (6) appropriate facies system. The two isopach sets do not represent the same ages, as is indicated by each map title. Abbreviated sequence names: Marsh, Marshalltown; U. English, Upper Englishtown; Merch, Merchantville; Cheese, Cheesequake; M, Magothy; and BR, Bass River.

paleogeographic maps and highstand sandbody isopach maps (Fig. 4) were integrated with total sequence isopach maps (Figs. 6–9) and observations from the offshore New Jersey margin (Poag and Sevon 1989) to construct a comprehensive record of New Jersey margin deposition, depocenter migration, and Late Cretaceous deltaic evolution.

This depositional history records the long-term signal of coastal onlap characteristic of post-rift passive margins (Grow 1980), apparent in the transition from the fully terrestrial and fluvial Albian Potomac Formation (deposited prior to the Bass River sequences) to the strong marine influence on deltaic facies during the Late Cretaceous (Miller et al. 2004). Thermoflexural subsidence is modulated by higher-frequency variations in sediment supply, flexural subsidence from sediment loading of the shelf, and third-order eustatic changes. Analysis of paleogeography and depocenter migration reveals five primary phases of Late Cretaceous margin deposition:

(1) Cenomanian–early Turonian Bass River I–III sequences exhibit the first evidence of marine strata on the coastal plain and the onset of 35 Myr of Late Cretaceous deltaic margin sedimentation. The paleoshoreline was oriented slightly more NE–SW than modern trends, representing the disparity of rapidly prograding delta fronts versus slower progradation of southern shoreface deposits (Fig. 4).

Deposition of the Bass River I and II sequences saw northern and central fluvial systems supply a broad delta front in the central coastal plain. These sands: (1) transition into prodelta and thin glauconitic shelf sands farther offshore; and (2) grade alongshore into thinner shoreface sands to the southwest, likely representing a transition to the bordering interdeltaic margin (Fig. 4). While Bass River I and II also record extensive delta-plain and fluvial sediments (supported by outcrop studies of the time-equivalent and updip Raritan Formation; Owens and Gohn 1985), Bass River III deposition is characterized by significant shoreline retreat (the result of higher sea level; Miller et al. 2005) and consists of thick, well-defined marine delta-front, prodelta, and shelf facies.

Sequence and sand depocenters are concentrated along the south-central coastal plain and are relatively stable through the Cenomanian–early Turonian, likely the result of a stable northern to central source and sediment supply. These results are consistent with offshore interpretations that identify a major depocenter off the coast of central New Jersey. A northern source appears to feed an offshore depocenter and bypass the northern coastal plain, resulting in slightly thinner intervals (Poag and Sevon 1989) (Fig. 10).

(2) A major northeast shift in depocenter location occurs during the late Turonian–Coniacian (Magothy sequences) associated with a long-term phase of low sea level (Miller et al. 2005), the establishment of two sediment delivery systems (northern and southern sources), a significant influx of sediment to the mid-Atlantic Margin (Poag and Sevon 1989), and increased subsidence in the Raritan Embayment (Fig. 10). These sequences thicken substantially offshore toward the Long Island Platform, where sections exceed 350 m (Fig. 10) (Poag and Sevon 1989).

Late Turonian–Coniacian deltaic sequences exhibit a wide array of marginal to nonmarine facies that are unique to Late Cretaceous deltaic sedimentation. These include delta and alluvial plain, paleosols, fluvial channel, levee, crevasse splay, lagoon, swamp, estuarine, and tidal channel to delta facies observed in core and outcrop (Miller et al. 1998b; Miller et al. 1999; Miller et al. 2006; Sugarman et al. 2005).

Paleogeographic analysis reveals several interesting trends of Turonian–Coniacian deltaic sedimentation: (1) The Magothy I sequence consists of thick delta-front deposits, but also exhibits delta-plain and fluvial and estuarine deposits across the southern and northern coastal plain. (2) No marine facies are recorded in the Magothy II sequence due to substantial progradation of the shoreline. Two extensive fluvial systems are visible in the northern and southern coastal plain, while thin alluvial and delta plain paleosols are abundant throughout (Fig. 4). (3) The Magothy III sequence records the highest diversity of facies observed in core, and consists of a substantial delta front with abundant fluvial, crevasse-splay, lagoon, and swamp deposits across much of the coastal plain. (4) The Magothy IVA and IVB sequences are preserved only in the northern coastal plain, and record extensive nonmarine delta-plain, fluvial-channel, and coastal lagoon to swamp facies that border thin delta-front sands to the east (Fig. 4). The restricted distribution of these northern sequences represents the dominance of a strong northern source and ample accommodation in the Raritan Embayment.

(3) Santonian deposition is characterized by a sediment-starved glauconitic shelf (Miller et al. 2004). No significant depocenters or deltaically influenced facies are visible on the coastal plain, representing a major transgression that caused relatively high sea level and a significant reduction of local siliciclastic input (Fig. 10).

(4) The Campanian is characterized by thick northern depocenters, although several secondary central depocenters are also evident. This is consistent with offshore intervals that record thick sections east of northern New Jersey (Fig. 10) (Poag and Sevon 1989). Campanian trends indicate the influence of both sediment sources: (1) the southern source was the primary control of a large tide- and wave-dominated delta (Merchantville sequence) as evidenced by a major depocenter on the south-central coastal plain; (2) the northern source deposited thick delta-front sands across the northern coastal plain (upper Englishtown sequence); and (3) significant northern and central depocenters of the Marshalltown sequence indicate both sediment sources were significant contributors to deposition (Fig. 10).

Delta-plain deposits are absent during the Campanian, and sequences exhibit a strong marine influence with significant accumulations of delta-front sands, prodelta sandy, silty clays, and glauconitic shelf sands (Fig. 4). Campanian sequences also document the rapid lateral transition from thick delta-front to shoreface facies, implicating significant wave reworking and longshore transport of this mixed-influence Cretaceous delta (Fig. 4).

(5) Maastrichtian deposition exhibits gradual thickening to the north and south while offshore maps exhibit broad contours that extend gently across the shelf towards a depocenter located $\sim$ 300 km to the east. The relatively thin and sediment-starved Maastrichtian Navesink sequence exhibits little influence from either source due to low sedimentation rates and deep paleodepths (middle to outer neritic; Miller et al. 2003) tied to high sea level (Miller et al. 2005). This period of deposition is unique because it: (1) lacks primary onshore depocenters; (2) is the only upper Cretaceous sequence to thicken toward the southern coastal plain; and (3) exhibits shelf facies with little to no deltaic influence from either sediment source.

The paleogeographic distribution of sequences reveals that shifting northern and southern sediment sources fed large deltaic systems and onshore and offshore depocenters. A northward shift in deposition from the Cenomanian to the Campanian resulted from a dominant northern source, a weakened southern source, and persistent thickening into the Raritan Embayment. The progressive shift from marginal and nonmarine deltaic facies in the Turonian–Coniacian to fully marine deltaic facies in the Campanian represents continued thermoflexural subsidence and a long-term rise in sea level.

Controls on the Distribution of Sequences and Facies

Late Cretaceous sequences and deltaic facies systems of the New Jersey Coastal Plain reflect the interplay of several variables: (1) eustatic variations dominate the timing of sequences, systems tracts, and generation of bounding disconformities; (2) differential flexural subsidence of the continental crust across the margin provides excess accommodation in the Raritan Embayment relative to the South Jersey High; (3) changes in tectonic uplift and weathering of Appalachian source terrains affects the rate and location (e.g., dominant fluvial axes) of sediment supply, influences the expression and characteristics of deltaic facies, and may have a positive feedback on the basement response to sediment loading.

Sea-Level Changes.—Third-order sea-level changes are well documented for the Late Cretaceous (Miller et al. 2005). One-dimensional backstripping estimates from New Jersey Coastal Plain coreholes identified 11 (and as many as 18) sea-level cycles from 100 to 65 Ma with amplitudes as great as $\sim$ 50 m (Miller et al. 2005). These sea-level changes are the principal driver behind base-level changes, unconformity genesis, and the timing of transgressions, regressions, and systems tracts on the New Jersey Coastal Plain.

Because thermoflexural subsidence is the dominant tectonic component of evolution of passive margins, New Jersey offers an excellent location to examine the evolution of eustatically forced sequences and deltaic facies. Periods of elevated or low sea level have a distinct effect on shoreline position and the types of deltaic facies that are recorded on the coastal plain. High sea level in the Campanian resulted in marine deltaic facies, while low Turonian–Coniacian sea level resulted in the deposition of marginal to nonmarine deltaic facies. However, eustasy alone does not account for the variability of deltaic facies across the coastal plain.

Our results from the Late Cretaceous show that although eustasy provides the template for sequences globally, regional tectonics (rates of subsidence and accommodation), autogenic changes in sediment supply, proximity to sediment input, and local subsidence from depocenter loading determines the preservation of sequences in a particular region.

Sediment Supply and Source Location.—The integration of Cenomanian–Maastrichtian paleogeographic and isopach maps of New Jersey sequences establishes a chronology of depocenter migration and documents the importance of two dominant sediment sources (northern and southern) on the distribution of deltaic sequences and facies. Changes in deltaic facies patterns and sandbody character appear to result from: (1) variations in sediment supply; (2) changes in source location and/or the dominance of a particular source; (3) proximity of sandbodies to a sediment source; and (4) the modifying effects of wave and tidal energy on deltaic facies distributions. These changes in sediment source location and sediment yield are superimposed on longer trends of basement subsidence and third-order eustatic variations.

For this study, the terms "northern" and "southern" source replace the "ancient-Hudson" and "ancient-Delaware" of Poag and Sevon (1989), who used the names to represent inferred locations of Appalachian drainage. Although fault patterns in Paleozoic basement may constrain the location of the modern Hudson River into the Cretaceous (lending a degree of permanence to its impact on New Jersey margin deposition), little work has addressed the issue and its precise location remains uncertain before the Plio-Pleistocene (Stanford et al. 2001). Similarly, the geological context and Cretaceous path of the modern Delaware River is unknown, although a southern source was likely located around the central coastal plain, much farther north than the modern Delaware River. Poag and Sevon (1989) infer the Adirondack Highlands as the primary source of ancient-Hudson sediment, with some influence from the western New England Highlands, while the ancient-Delaware is fed by the Central Appalachian Highlands (Fig. 1).

Long-term trends reveal the gradual shift of depocenters from the central to northern New Jersey Coastal Plain from the Cenomanian (ca. 98 Ma) to the Campanian (ca. 72 Ma). This reflects the nature of a two-sediment-source system and variations in sediment supply tied to extrabasinal uplift and increased weathering of source terrains (Poag and Sevon 1989). Peak rates (21 km^3/Myr) of Late Cretaceous sediment accumulation on the mid-Atlantic Margin occurred during the Coniacian, representing a phase of tectonic uplift and intense weathering of the Appalachian hinterland (Poag and Sevon 1989). This large influx of sediments is reflected by the rapid seaward progradation of the shoreline and preservation of extensive delta-plain deposits on the New Jersey Coastal Plain (Magothy sequences). The concentration of Magothy depocenters in the northern coastal plain implies a higher sediment load in the northern source than the southern source. Observations from the coastal plain are consistent with offshore data that shows large amounts of coarse, deltaic material deposited across New Jersey shelf, a function of high sediment rates "flooding" the system (Poag and Sevon 1989).

Conversely, periods of low to dormant uplift and weathering are characterized by a reduction in the amount of sediment delivered to the coast by the fluvial systems. Deposition during these intervals can be characterized by a retreat of the shoreline and the onset of largely sediment-starved, glauconitic shelves across the New Jersey Coastal Plain (although sea level also plays an important role). While the Maastrichtian exhibits substantial sediment accumulation rates (11 km^3/Myr) across the mid-Atlantic Margin, most of this sediment is derived from sources to the north and south of New Jersey (Poag and Sevon 1989). As a result, there is little evidence of any deltaic influence in New Jersey during these time periods, with the only shallow sands identified as distal lower shoreface in the Sea Girt corehole (Sugarman et al. 2005). Although the Cenomanian experiences extremely low regional sediment accumulation rates (2 km^3/Myr), the thickest intervals are located 100 km east of New Jersey. As a result, relatively thin but well-defined deltaic sequences are preserved across the New Jersey Coastal Plain while coeval sections are thin to absent across much of the mid-Atlantic Coastal Plain (Poag and Sevon 1989).

Proximity to fluvial axes and active deltaic lobes plays an important role in sequence thickness and the character (e.g., lithology, grain size, porosity, permeability) of deltaic sandbodies. Several sequences (Cenomanian and Campanian) exhibit thick delta-front sands that grade laterally into thinner shoreface sands over relatively short distances (5–50 km). While several of these could simply represent interdeltaic zones of the margin ("shore-zones" of Galloway 2001), it appears that large amounts of deltaically derived sand are reworked by wave action and redistributed by longshore currents. Such lateral variations in facies and sandbody character are important to understand, particularly in the application of ancient deltaic systems to hydrocarbon and hydrogeologic studies. The gradation from thick, porous delta-front sand to thinner, finer-grained lower-shoreface sands observed throughout upper Cretaceous sequences can significantly alter the viability of reserves. In non-hydrocarbon-bearing regions, these sandy intervals are also important aquifers, particularly in densely populated areas such as the greater New York–New Jersey–Philadelphia metropolitan area. Understanding the process and scale of such changes can be critical in effectively managing groundwater resources.

While the marine delta front is generally the locus of sand deposition of the Late Cretaceous New Jersey delta, the associated progradation of the delta often preserves an extensive delta plain where many nonmarine facies have significant quantities of sand. Most of the thick, coarse sands are found in fluvial to tidal channels that dissect the ancient delta plain. However, additional sandbodies can be uncovered in crevasse-splay, levee, lagoon, swamp, and bay deposits, although the lateral continuity of these sandbodies is limited and difficult to constrain with sporadic core and well coverage. Many of these sandy intervals are heterolithic accumulations of sand with paleosols and floodplain muds and clays, limiting their utility for hydrologic purposes on the coastal plain.

Basement Structure and Subsidence.—While thermal subsidence and subsequent flexural loading is the dominant form of subsidence on passive margins (Watts and Steckler 1979; Watts 1982; Kominz et al. 1998), a series of basement embayments and arches influence the structural fabric of the Atlantic Coastal Plain. This trend is manifest in New Jersey with the southern Salisbury Embayment, the smaller northern

Raritan Embayment, and the intervening South Jersey High (Fig. 1) (Olsson et al. 1988).

Isopach maps of upper Cretaceous sequences reveal the ubiquitous influence of the Raritan Embayment and South Jersey High on sequence distribution. Three Cenomanian–lower Turonian Bass River sequences exhibit thinning onto the South Jersey High (Fig. 10). Five Turonian–Coniacian Magothy sequences and three Campanian sequences thicken into the northern Raritan Embayment and similarly thin onto the South Jersey High (Fig. 10). Although they have influenced deposition since the Early Cretaceous, the genesis and behavior of these structural features is unclear (Olsson et al. 1988).

Brown et al. (1972) defined the tectonic framework of the coastal plain as a regional system of crustal segments that formed fault-bounded grabens as a result of far-field compression on wrench-fault zones. Differential subsidence along these fault-bounded "segments" was thought to deposit thick sedimentary sections in these embayments versus bordering basement highs. However, the existence of these large faults is unclear. Although active faulting has been observed across the Atlantic Coastal Plain south of the Salisbury Embayment (at Charleston, South Carolina; Weems and Lewis 2002), the New Jersey Coastal Plain shows no evidence of syndepositional faulting (Kominz et al. 1998), though antecedent faults such as the Cornwall–Kelvin fault under the Raritan Embayment (Drake and Woodward 1963) and a southern fault trending under Cape May (Taylor et al. 1968) could have provided inherited basement structure that influenced sequences (Browning et al. 2006). However, isopach mapping reveals no direct evidence of significant faulting during the Late Cretaceous such as large (+ 15 m) thickness variations over short distances (~ 2 km), growth packages, or erratic contours.

Variations in sequence thickness in the New Jersey Coastal Plain appear to result from "normal" passive-margin thermoflexural subsidence (Watts and Steckler 1979; Kominz et al. 1998) and the consequent flexural response of progressive point loading of thick sedimentary packages into the Raritan Embayment and farther offshore by a northern sediment source. This loading caused positive feedback and increased flexural subsidence that accentuated the existing basement fabric and increased accommodation rates for subsequent units. The thinning of units onto the South Jersey High may represent a peripheral bulge (e.g., Galloway 1989) caused by the progressive flexural response to Early Cretaceous and subsequent loading in the Salisbury and Raritan embayments.

Isopach mapping reveals trends that validate thermoflexural subsidence as the primary control of regional accommodation and sedimentation, namely the persistent thickening into the Raritan Embayment, thinning onto the South Jersey High, and broad continuous contours that extend across the coastal plain (Fig. 10). Similar to the work of Galloway (1989; Galloway et al. 2000) and Browning et al. (2006), we find that the position of embayments and structural highs can be largely attributed to syndepositional flexural subsidence due to large prograding sedimentary wedges across the shelf.

CONCLUSIONS

We use core- and geophysical-log correlation to map Upper Cretaceous sequences and deltaic facies across the New Jersey Coastal Plain and evaluate and refine well-log predictions in the absence of core control. Core-log correlations from four continuously cored ODP sites (Ancora, Bass River, Millville, and Sea Girt) establish a clear link between the identified sequences (based on lithology, biostratigraphy, and Sr-isotope dating) and their respective gamma-ray and resistivity geophysical log signatures.

Paleogeographic, isopach, and deltaic lithofacies mapping of thirteen depositional sequences established a 35 million year, high-resolution (> 1 Myr) record of Late Cretaceous deltaic sedimentation of the New Jersey Coastal Plain. Our study illustrates the widely known variability of deltaic systems, but also documents the relative stability of deltaic facies systems on the 10^6–10^7 yr scale, with long periods of cyclically repeating systems tracts controlled by eustasy.

This study reveals five primary phases of margin evolution during the Late Cretaceous: (1) Cenomanian–early Turonian deltaic facies shift from delta plain to fully marine and are thickest in the central coastal plain; (2) high sediment rates, low sea level, and high accommodation rates in the northern coastal plain resulted in thick, marginal to nonmarine mixed-influenced deltaic facies during the Turonian–Coniacian; (3) comparatively low sediment rates and high sea level during the Santonian resulted in a sediment-starved margin without clear deltaic influence; (4) Campanian deltaic sequences thicken to the north and exhibit wave reworking and longshore transport of sands; and (5) low sedimentation rates and high long-term sea level during the Maastrichtian resulted in a sediment-starved glauconitic shelf.

Deltaic facies characteristics are strongly influenced by long-term eustatic changes, allogenic variations in sediment supply, and proximity to two long-lived fluvial axes. Sequence depocenters migrate gradually northeastward from the Cenomanian (ca. 98 Ma) through the earliest Danian (ca. 64 Ma) and reflect the position of active sediment sources and flexural subsidence due to large prograding sediment loads on the coastal plain and offshore shelf.

Results from the Late Cretaceous show that although eustasy provides the template for sequences globally, regional tectonics (rates of subsidence and accommodation), autogenic changes in sediment supply, proximity to sediment input, and flexural subsidence from depocenter loading determines the regional to local preservation and facies expression of sequences.

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