Holocene evolution of an estuarine coast and tidal wetlands

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ABSTRACT

Modern facies-distribution patterns, extensive core data, and chronostratigraphic cross sections provide a detailed history of Holocene inundation within the Delaware Bay estuary and sedimentation in adjacent coastal environments. Flooding of the estuary occurred with rising sea level as the shoreline retreated northwest along a path determined by the pre-transgression topography. Simultaneous migration of an estuarine turbidity maximum depocenter provided the bulk of fine sediments which form the coastal Holocene section of the estuary. Prior to 10 Ka, the ancestral bay was predominantly a tidal river, and the turbidity maximum depocenter was located southeast of the modern bay mouth. By 10 Ka, lowlands adjacent to the ancestral channel of the Delaware River were flooded, forming localized tidal wetlands, and the depocenter had initiated high rates of fine-grained sedimentation near the present bay mouth. At that time, coastal Holocene strata began to onlap the interfluve highlands. By 8 Ka, the fine-grained depocenter had migrated northwest along the main channel of the Delaware River, although the widened mouths of tributary valleys continued to be active sites of sediment accumulation. Following the passage of the fine-grained depocenter, coarse-grained sediments accumulated along the coast in response to increased wind-wave activity. During the middle Holocene, portions of the estuarine coast began to resemble modern geomorphology, and washover barrier sands and headland beach sandy gravels accumulated along the southwest shore. The late Holocene was characterized by erosional truncation and submergence of aggraded coastal lithofacies and by planation of remnant highland areas. Knowledge of the eroded Holocene section is fragmentary. At present, continued sea-level rise is accompanied by deposition of tidally transported muds in coastal environments and deposition of sandy sediments in some offshore regions. An unconformity marks the base of the developing open estuarine sequence of coarse clastic lithofacies and denotes the end of coastal accumulations. Modeling of coastal-lithofacies transitions identifies specific lithofacies complexes in the Holocene stratigraphic section which were influential in the evolution of the coast. Development of the Holocene section of the estuary coast involved both constructive, or aggradational, and destructive, or erosional, phases.

INTRODUCTION

Recent work on the geologic history of the Holocene transgression in the region of Delaware Bay estuary (Figs. 1A, 1B, and 1D) by Belknap and Kraft (1985), Chrzastowski and Kraft (1985), Fletcher (1986), Fletcher and Kraft (1986a), Fletcher and Pizzuto (1986), Knebel and others (1988) (Fig. 1C), Knebel and Circé (1988), and Knebel (1989) have expanded on the earlier investigations of Moody and Van Reenan (1967), Moose (1973), Sheridan and others (1974), Kraft and others (1975), Weil and others (1975), Weil (1976, 1977), and Twichell and others (1977). From these newer studies, a more detailed picture of the evolution of Delaware Bay estuary is emerging.

In this paper, we describe the evolutionary history of the coastal sedimentary system along the southwest shore of Delaware Bay. Knowledge of the modern distribution of sedimentary facies (Fig. 1E) in both coastal and offshore environments (Jordan and Groot, 1962; Parker and others, 1964; Jordan, 1968; Kraft, 1971, 1979; Kraft and others, 1973, 1979; Owens and others, 1974; Richter, 1974; Sheridan and others, 1974; Weil, 1976, 1977; Maley, 1981; Marx, 1981; Fletcher, 1986), extensive core and seismic data (Sheridan and others, 1974; Richter, 1974; Weil, 1976, 19777; Maley, 1981; Marx, 1981; Fletcher, 1986), and paleogeographic reconstructions (Weil, 1976, 1977; Kraft and Belknap, 1986; Fletcher, 1986; Fletcher and Kraft, 1986a; Knebel and others, 1988), comprise the large data base for understanding the Holocene development of this coast. In addition, we present here four chronostratigraphic cross sections of the southwest estuarine coast which typify the stratigraphy of the estuarine margin. Our description includes a reconstruction of flooding patterns, the chronology of basin inundation, a model of the aggradation and subsequent erosion of the coastal Holocene stratigraphic section, and a synopsis of the evolutionary development of the estuarine coast.

DEPOSITIONAL FRAMEWORK

The Modern Estuary

Delaware Bay is one of the largest Coastal Plain estuaries on the United States Atlantic coast. Draining more than 32,000 sq km, the Delaware River becomes tidal for a length of 60 km near Trenton, New Jersey, where the head of tide is located. The upper tidal river is generally unaffected by marine salinity; however, a transitional zone, characterized by salt water and fresh water mixing, occurs between 60 and 140 km downstream (Cronin and others, 1962; Harleman and Ippen, 1967; Weil, 1976, 1977). Delaware Bay, classified as a partially mixed estuary along its narrow upper reach, extends for about 75 km from the lower tidal river, eventually becoming a well-mixed estuary where it broadens at Port Mahon, Delaware (Fig. 1D) (Weil, 1976, 1977; Schubel and Meade, 1977; Hull and others, 1986). Nontidal mixing drives a circulation system

This article is a contribution to International Geological Correlation Program Project No. 274: Coastal Evolution in the Quaternary. Additional material (Table 2) for this article may be secured free of charge by requesting Supplementary Data 9004 from the GSA Documents Secretary.

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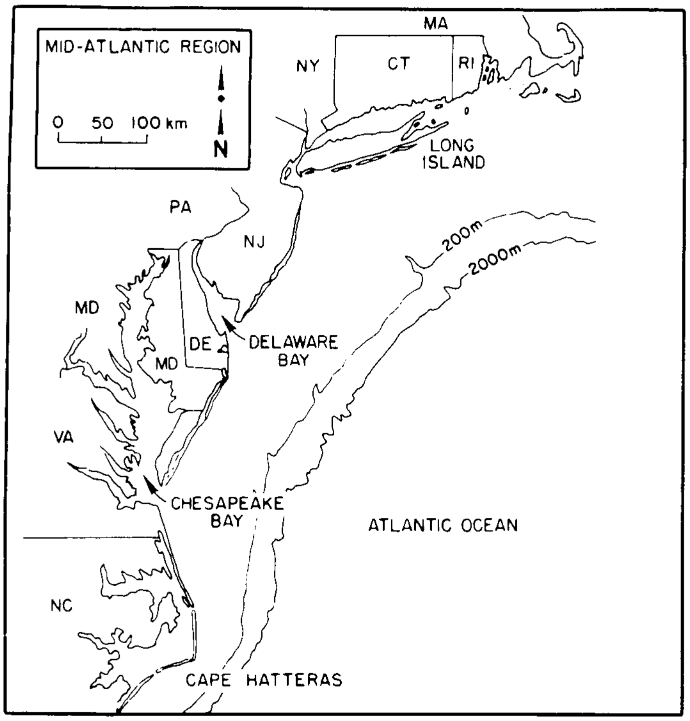

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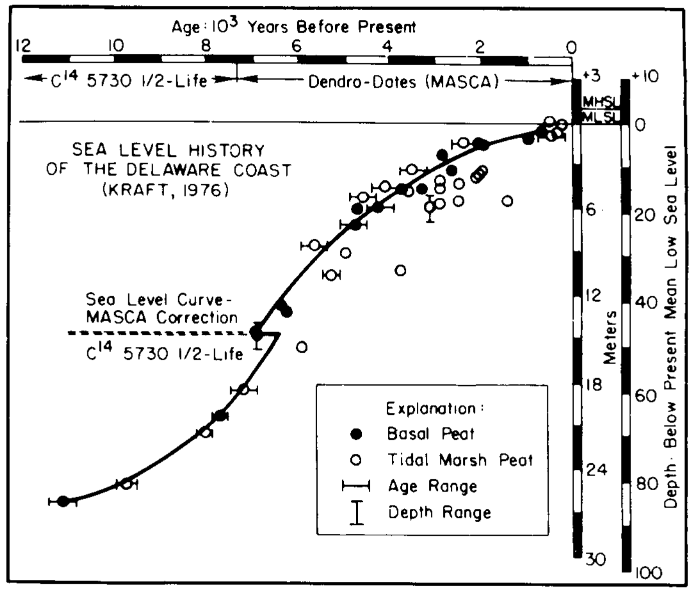

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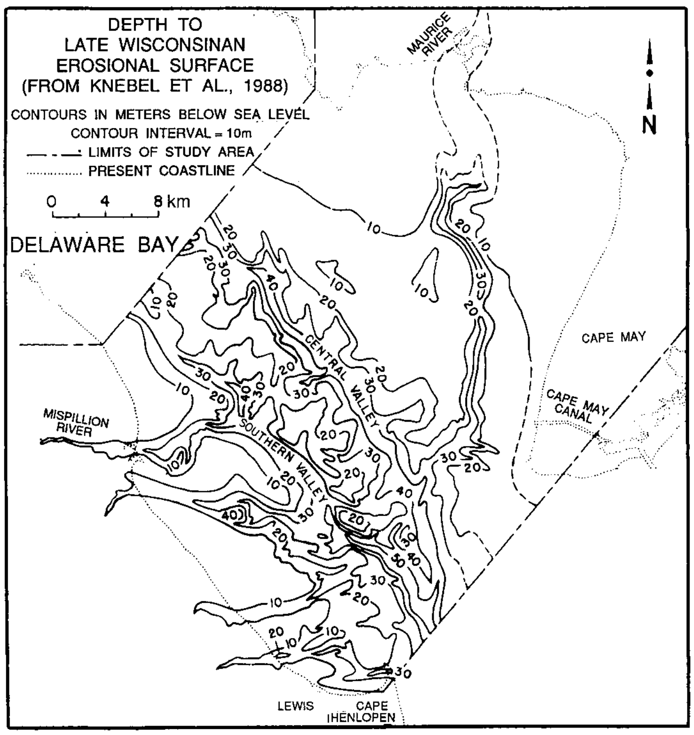

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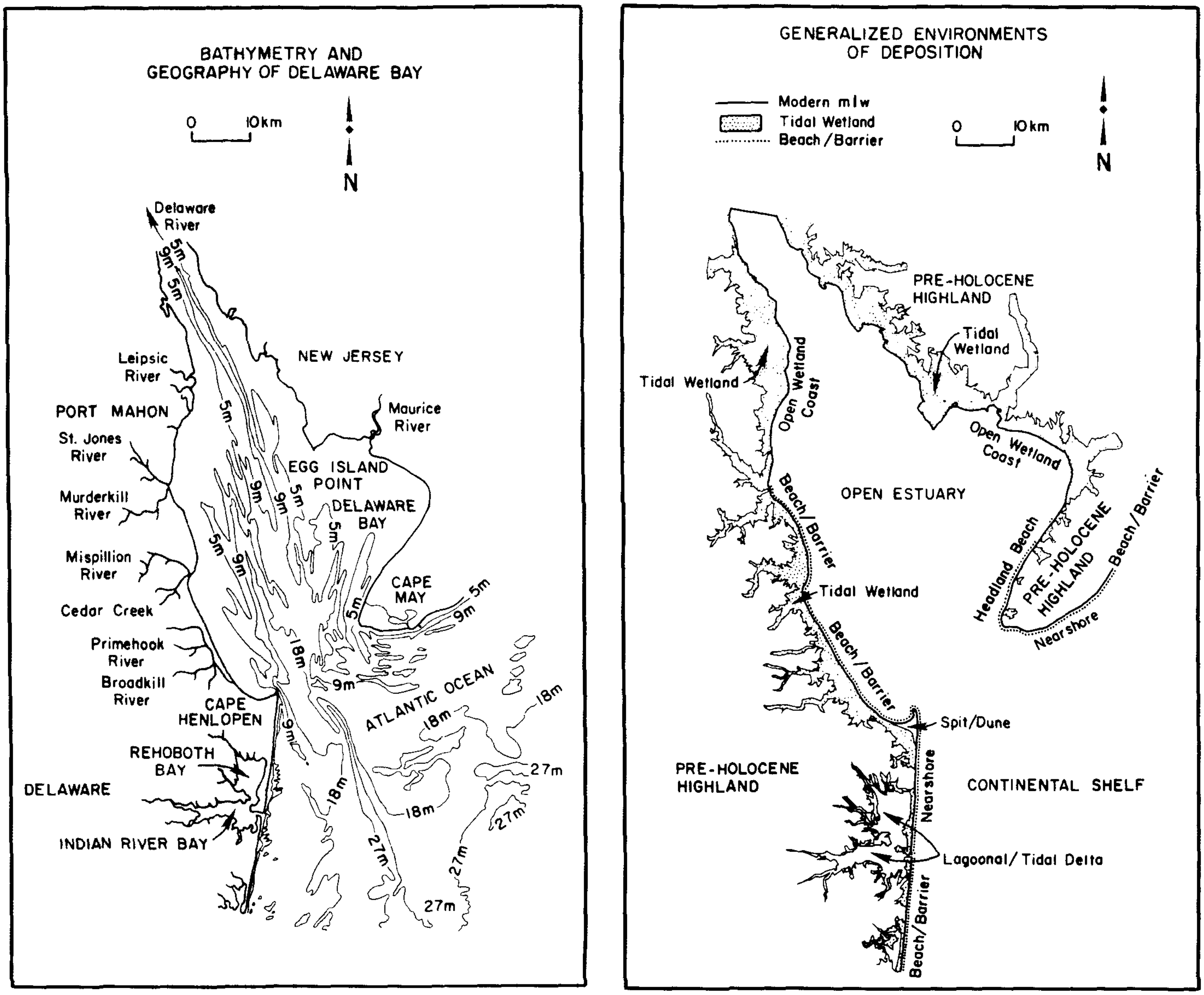
Figure 1. Composite illustrations of the Delaware Bay estuary setting, including (A) geography of the mid-Atlantic region; (B) Holocene local relative sea-level history of the Delaware coast from Kraft (1976); (C) pre-transgression topography of the central and southeast estuary, determined from seismic interpretation (Knebel and others, 1988).

in the lower tidal river which leads to formation of a turbidity maximum, characterized by rapid accumulation of fine-grained sediments (Harleman and Ippen, 1967; Weil, 1976, 1977; Gibbs and others, 1983; Oostdam, 1985). The open estuary is characterized by complex bathymetry produced by tidal current activity, whereas wind-wave related processes operate at the estuarine margin (Sheridan and others, 1974; Weil and others, 1975; Weil, 1976, 1977) (Figs. 1D and 2). Landward of the estuary shore, flooding of the paleodrainage system (Figs. 1C and 3) and deposition of fine-grained sediments has produced a complex of tidal wetlands (Kraft, 1971, 1979; Sheridan and others, 1974; Richter, 1974; Weil and others, 1975; Weil, 1976, 1977; Field, 1979).

Lithofacies Characteristics

A number of lithofacies variants (Table 1) are found in core sections along the estuary coast. Although these sediments are representative of a complex system of depositional environments, they may be characterized as belonging to four major lithofacies groups: the pre-transgression lithofacies, the tidal wetland lithofacies, the barrier-spit lithofacies, and the open-water lithofacies.

The Pre-Transgression Lithofacies Group. This group is composed of basal Holocene and pre-Holocene sediments. Basal Holocene sediments, typically early Holocene in age, consist of gravelly and silty sands of fluvial and fluvio-tidal origin that were deposited along pre-Holocene valley axes prior to the rise of sea level (Kraft, 1971, 1976). These deposits may reach thicknesses of several meters (Jordan and Groot, 1962; Owens and others, 1974; Talley, 1985). Pre-Holocene lithofacies are typically well-oxidized coarse clastics of either Miocene or late Pleistocene age (Jordan, 1964, 1974; Delaware Geol. Survey, 1976; Owens and Denny, 1979; Belknap, 1979; U.S. Geol. Survey, 1987). In some places laminated with thin mud beds, the sediments characterizing this lithofacies are in most cases well oxidized, gravelly, poorly sorted sands (Jordan, 1964, 1974; Owens and Denny, 1979). Sedimentary structures include cross-bedding, desiccation



D E

Figure 1. ***(Continued).*** **(D) Bathymetry and geography of the estuarine basin and continental shelf; (E) generalized environments of deposition in the estuary, surrounding shorelines, and along the Delaware Atlantic coast.**

features, mottled coloring, and burrows. Detrital plant and shell fragments are common (Richter, 1974; Kraft and John, 1976; Maley, 1981; Marx, 1981; Fletcher, 1986). Flooring some of the tributary channels, and found under the late Pleistocene deposits of the Atlantic coast, a sandy to gravelly white clay has been described which may be stiff, fluidized, or chalky and often has a bluish tint (Demarest, 1981; Fletcher, 1986). The topography of the pre-Holocene surface is a result of superimposed late Illinoian and late Wisconsinan drainage (Knebel and Circé, 1988) and has been shown to control the inundation patterns of the Holocene transgression (Belknap and Kraft, 1981, 1985).

The Tidal-Wetland Lithofacies Group. This group includes high and low salt marsh often found forming a basal peat in cores of the coastal section. Fresh-water tidal deposits from riverine and palustrine systems (including fresh-water peats) may also be found in basal facies (Whallon, 1989). Where marsh (or palustrine) lithofacies cannot be distinguished, the general term "tidal wetland lithofacies" is used; this consists of possible marsh, lagoon (if present), tidal-stream, and tidal-mudflat sediments. Finkelstein and Hardaway (1988) recognized cyclical stages of marsh and estuarine sedimentation in the fringing marshes of the York River, Virginia. Although a stratigraphy of this type has not been proposed for Delaware Bay, the sea-level history of the two areas is similar. The landward fringe of the Delaware Bay estuarine basin is characterized by salt marsh with floral zones indigenous to specific tidal flooding cycles (Parker and others, 1964; Richter, 1974; Kraft and others, 1975; Allen, 1977). Salt marsh also populates the landward surface of washover barriers between depositional events. Salt-marsh sediments are typically brown mud, or sandy mud with organic detritus, or in-growth-position plant fragments; peat layers are common (Kraft, 1971; Kraft and others, 1973; Richter, 1974; Weil, 1976, 1977; Allen, 1977). What have in the past been interpreted as lagoonal sediments are typically gray-brown muds with inclu-

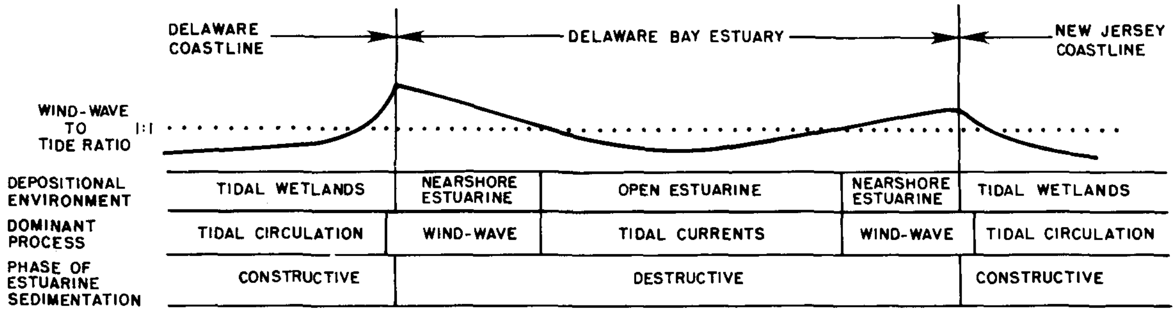


Figure 2. General description of depositional environment, dominant process, and phase of estuarine sedimentation across the Delaware Bay (from C. B. Weil, 1989, personal commun.).

sions of plant and shell debris, and some sand and gravel laminae (Kraft, 1971; John, 1977; Weil, 1976, 1977; Maley, 1981; Marx, 1981). The past occurrence of lagoonal environments along the coast of Delaware Bay, however, has been challenged in the paleogeographic reconstructions of Chrzastowski and Kraft (1985), Chrzastowski (1986), Fletcher (1986), Knebel and others (1988), Khalequzzaman (1989), and Whallon (1989). Church and others (1987) attributed the accumulation of the salt-marsh component of tidal-wetland sediments to deposition of estuarine seston by rare high-intensity storm events occurring a few times a year or less. The work of Finkelstein and Hardaway (1988), Jordan and Correl (1986) and Stumpf (1981) also emphasized the importance of storm-related deposition of suspended estuarine sediments on the salt-marsh surface.

The Barrier-Spit Lithofacies Group. Gravelly sands and sands of the washover barriers and headland beaches of the southwest coast and the prograding spit system at Cape Henlopen make up this group. The continuous sandy shoreline of the southwest coast of Delaware Bay is the product of incident wind waves generating near-bottom orbital velocities sufficient to entrain sandy and gravelly sediments and to suspend finer material (Weil, 1977; Coughanowr, 1985). The inundation of the estuary, over the course of the transgression, brought open water of sufficient fetch to produce waves capable of building sandy barriers, eroding headlands,

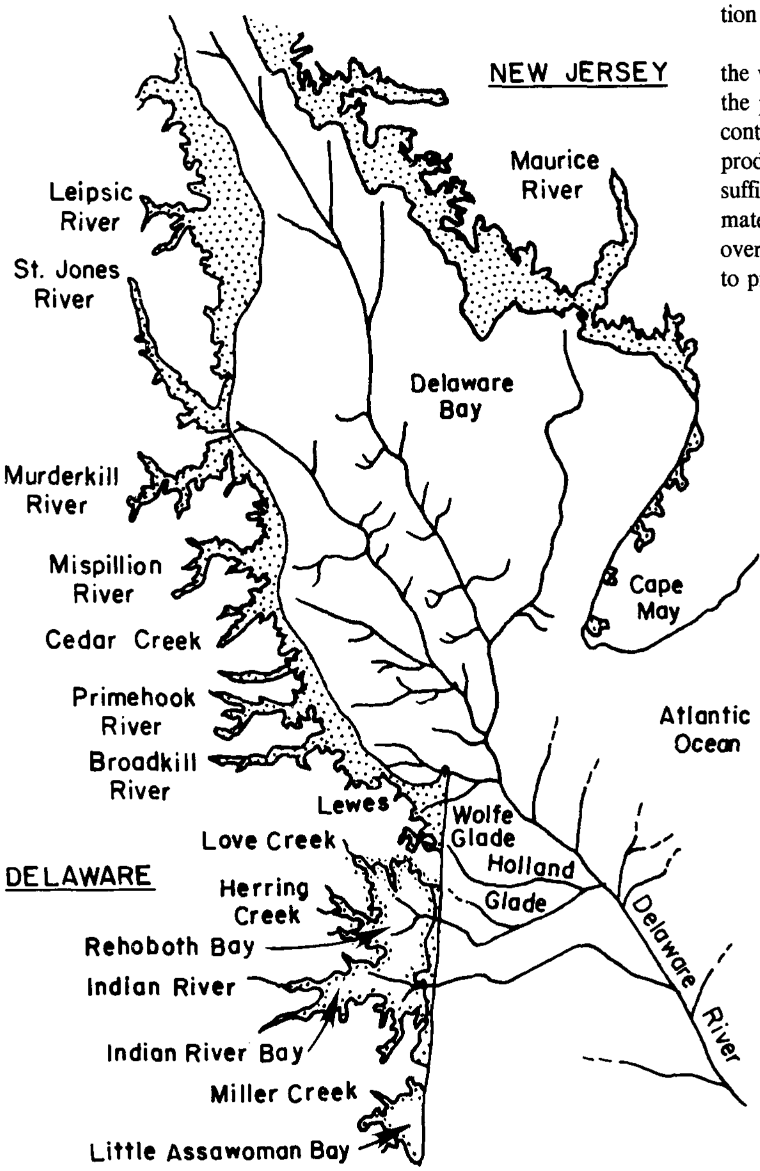


Figure 3. Paleogeographic reconstruction of the late Wisconsinan drainage system of the Delaware Bay (compiled from Richter, 1974; Sheridan and others, 1974; Weil, 1976; Field, 1979; Kraft and Belknap, 1986; Knebel and Circé, 1988; Knebel and others, 1988). Stippled pattern indicates extent of modern tidal wetland.

TABLE 1. DESCRIPTION OF LITHOFACIES GROUPS AND CROSS-SECTION CODES

Pre-transgression	
Basal Holocene	
Gravelly, silty sand; poorly sorted, olive gray to yellow	A1
Gravelly sand; poorly sorted	A2
Silty sand; poorly sorted	A3
Pre-Holocene	
Gravelly sand; poorly sorted, mud laminae	P1
Gravelly sand; poorly sorted, cross-beds, oxidized	P2
Sand; poorly sorted, compacted, oxidized	P3
Mud; interbeds of sand and gravel, oxidized, desiccated	P4
Sandy mud/muddy sand; cross-beds, plant fragments, oxidized	P5
Sandy, gravelly clay; white, stiff or fluidized, chalky with bluish tint, oxidized	P6
Sandy, gravelly mud; cross-beds, plant/shell fragments, oxidized	P7
Mud; fluidized or compacted, burrowed or laminated, oxidized	P8
Tidal wetland	
Marsh (where distinct)	
Mud; *in situ* plant material, some shell fragments	M1
Peat	M2
Sandy mud; plant fragments, gravel rare	M3
Tidal wetland (tidal stream, high and low marsh, tidal mud flat)	
Mud; gray brown, plant fragments, peaty layers, occas. sand, gravel, and shells	W1
Barrier – spit	
Washover barrier and headland beach	
Gravelly, silty sand; may contain laminations, shell or plant detritus, cobbles, poorly sorted	B1
Sand; poorly sorted, may contain shell or plant detritus, some heavy minerals	B2
Gravelly sand; variable sorting, may contain organic detritus	B3
Silty sand; variable sorting, may contain organic detritus	B4
Prograding spit	
Gravelly sand; variable sorting, shell fragments, peat rollers, clay balls; swash-zone debris common	S1
Sand; poorly sorted, organic detritus	S2
Gravelly, muddy sand; poorly sorted, some laminations	S3
Open water	
Estuarine (non-specific)	
Clayey silt; may contain organic detritus	E1
Sandy mud; some shell and plant detritus, laminations or wormtubes or other bioturbational structures	E2
Gravelly, sandy mud	E3
Estuarine (shallow)	
Muddy, gravelly sand; poor sorting, some fine laminae, organic detritus	X1
Sand; variable sorting, laminated or bioturbated, organic detritus	X2
Mud to coarse sand; finer sizes laminated, some shell or plant fragments, graded beds, burrows, mottling, shell hash	X3
Similar to X3 with absence of mud	X4
Fine to coarse sand; interpreted as under marine influence (shelf)	Y2

and transporting coarse clastics alongshore (Kraft, 1971; Weil, 1976, 1977; Maley, 1981; Marx, 1981; Fletcher, 1986; Knebel and others, 1988). Headland-beach sediments are typically fine to coarse sands with shell fragments and some gravel. Barriers consist of gravelly to silty sands of variable sorting with some heavy mineral or silt laminations, shell or plant detritus, and some cobbles (Weil, 1976, 1977; Kraft and others, 1973, 1979; Maley, 1981; Marx, 1981; Fletcher, 1986).

The Open-Water Lithofacies Group. This group consists of a variety of sediment types which are reflections of dominant sedimentary processes operating in the estuary. Coarse materials include the reworked and winnowed sands and gravels of the tidal channels (Weil and others, 1975; Weil, 1976, 1977; Knebel, 1989), the shoals and banks of the estuarine floor (Weil and others, 1975; Weil, 1976, 1977; Knebel, 1989), the longshore contribution from the southerly littoral drift system of the Cape May Atlantic coast, and the northerly longshore system of the Delaware Atlantic coast, as well as the sands transported by flooding tide from the continental shelf (Duane and others, 1972; Swift and others, 1972; Neiheisel, 1973; Sheridan and others, 1974; Weil, 1976, 1977). Fine-grained sediments consist mostly of clayey silt, sandy mud, or mud with organic detritus and some bioturbational structures (Sheridan and others, 1974; Weil, 1976, 1977). These muds have accumulated within channels along the axes of the ancestral Delaware River and associated tributary valleys due to the northwest migration of the turbidity maximum (Fletcher, 1986, 1988a).

STRATIGRAPHY

Assumptions and Presentation

Chronostratigraphic cross sections of the southwest estuarine coast are used in the following discussion. These are constructed along shore-normal and shore-parallel transects from a data base of 100 vibracores (Fig. 4) obtained in a number of studies (Richter, 1974; Sheridan and others, 1974; Kraft and John, 1976; Weil, 1976; Maley, 1981; Marx, 1981; Fletcher, 1986). Lithosome tops were age estimated using local relative sea-level history (Fletcher, 1988b); these are tabulated in Table 2.[1] The following is a description of the use of local relative sea-level history as a means of estimating lithosome ages.

Relative sea-level age estimation is an application of radiocarbon sea-level history which assumes that a curve constructed of carbon-14 dated basal salt-marsh peats accurately represents the elevation of past sea levels relative to modern mean low water (Kraft, 1976; van de Plassche, 1986; Fletcher, 1988b). The sea-level history described by the Kraft (1976) curve is translated into a series of numerical relationships (Fig. 5). These allow use of iterative techniques as a means of estimating the ages of the more than 1,000 separate lithosomes in the vibracore data base (Fletcher and Kraft, 1986b; Fletcher and Pizzuto, 1986; Fletcher, 1986). The technique of age estimation has a probable maximum error of ±500 yr based on an evaluation of the influence of compaction on the elevation of

[1]Table 2 may be secured free of charge by requesting Supplementary Data 9004 from the GSA Documents Secretary.

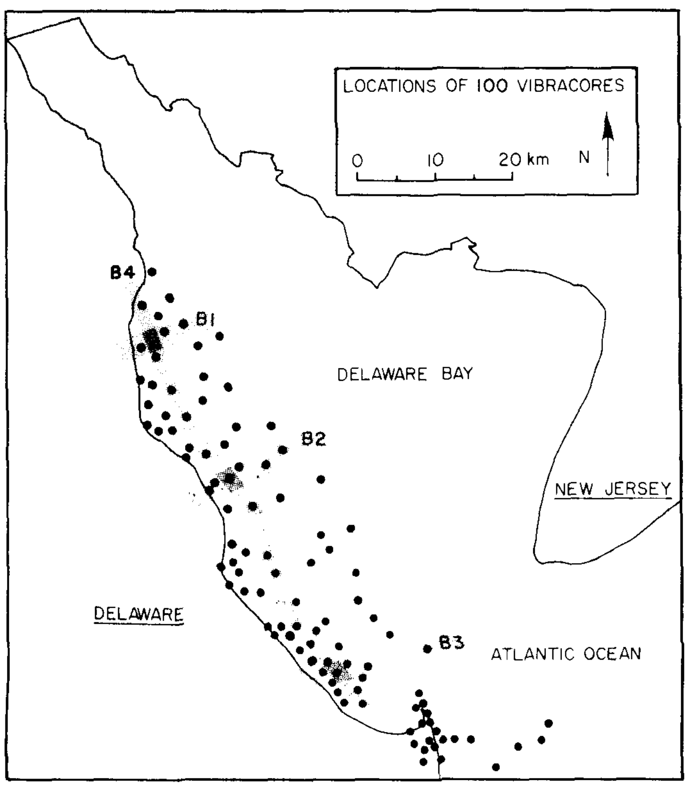


Figure 4. Location map of the vibracores used in this study (compiled from Richter, 1974; Sheridan and others, 1974; Kraft and John, 1976; Weil, 1976; Maley, 1981; Marx, 1981; Belknap and Kraft, 1985; Fletcher, 1986). Shaded transects are shown in cross section in Figures 7 and 8.

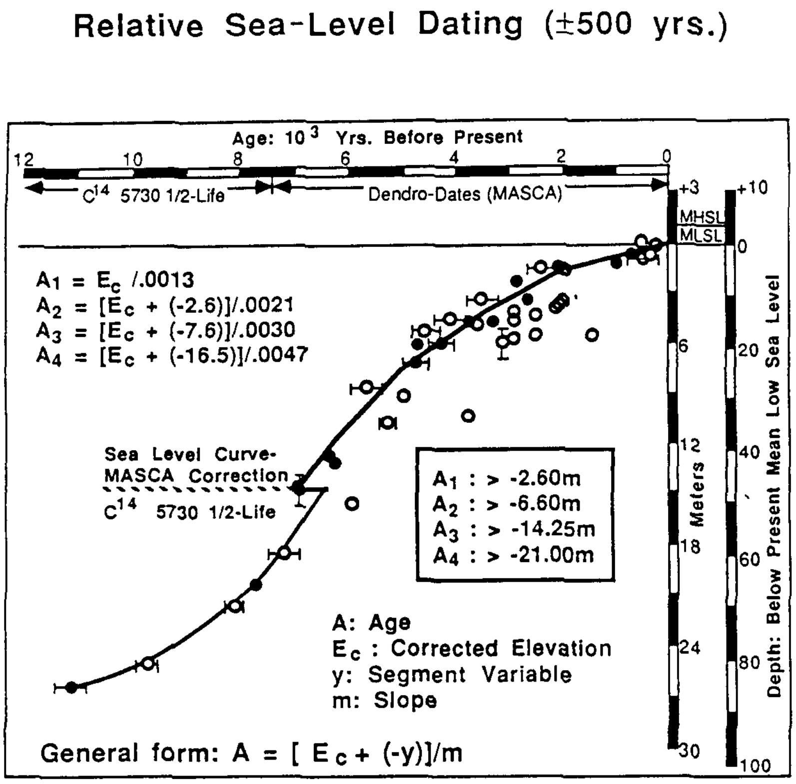


Figure 5. The Kraft (1976) history of local relative sea level is used in the technique of lithofacies age estimation (Fletcher, 1986; Fletcher and Pizzuto, 1986; Fletcher and Kraft, 1986b; Fletcher, 1988b). Four straight-line segments of the curve are used to predict the age (A) of a cored lithofacies using the deposit elevation (corrected for environment of deposition relative to paleomean low water, Ec), and the slope of the segment (m). The segment variable (y) identifies the appropriate equation for the deposit.

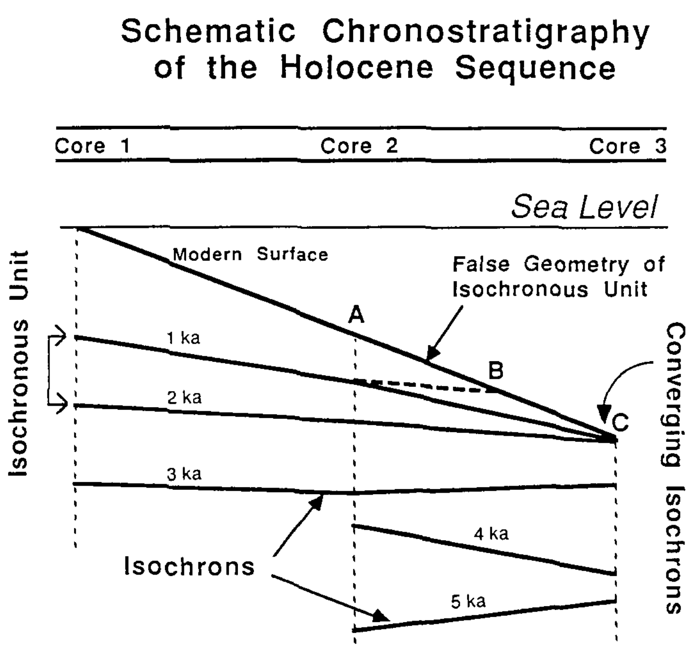


Figure 6. The chronostratigraphic cross sections of Figure 7 and 8 are plotted in a manner that suggests isochronous unit pinch-out, when in fact the seaward extent of most coastal isochronous units is truncated by erosion at the shoreface. This creates a false geometry. For instance, the unit 1 Ka to 2 Ka should have the dashed line as the seaward extent of its upper isochron and should include the volume of material underlying the segment BC. The constraints of the data base, however, require the use of a pinch-out depiction (Fletcher and Kraft, 1986b).

dates derived from nonbasal materials (Pizzuto and Fletcher, unpub. data; Fletcher, 1988b) and the influence of paleotidal range (Decker, 1984; Fletcher, 1988b) and local datum change (Holdahl and Morrison, 1974; Fletcher, 1988b).

The chronostratigraphic cross sections illustrate the aggradation history of the coastal Holocene section by depicting the occurrence of millenial isochronous units (Fig. 6). An isochronous unit is a continuous succession of strata bounded by synchronous surfaces representing an interval of time (Krumbein and Sloss, 1963). The synchronous surfaces are shown by isochrons correlating relative sea-level estimated ages of cored lithologies (Figs. 7 and 8, illustrated with an alphanumeric code given in Table 1).

The relationship and character of adjacent isochrons illustrate the geometry of successive isochronous units. Isochron placement is controlled by the density of correlatable age estimates. The constraints of the data base require cross-section graphics which fail to show the truncation of isochrons by the erosion of the shoreface with sea-level rise. Rather, a pinch-out is implied in the graphics (Fig. 6), which erroneously suggests that the transgression changes the geometry of isochronous units. Where an interval suggests that an isochron is pinching-out, the true stratal geometry involves truncation of the unit, not the convergence of isochrons.

Chronostratigraphy of the Coastal Section

In all of the chronostratigraphic cross sections (Figs. 7 and 8), the pre-transgression lithofacies is basement to the Holocene section. Due to the active drainage which incised the Delmarva Peninsula during the late Wisconsinan, much of this surface is lowland, formerly occupied by tributary channels to the ancestral Delaware River. These areas are readily flooded by rising sea level (Kraft, 1971; Kraft and John, 1976; Belknap and Kraft, 1981, 1985). Incised in this manner, the pre-transgression surface provides for the accumulation of a thick sedimentary on-lap sequence

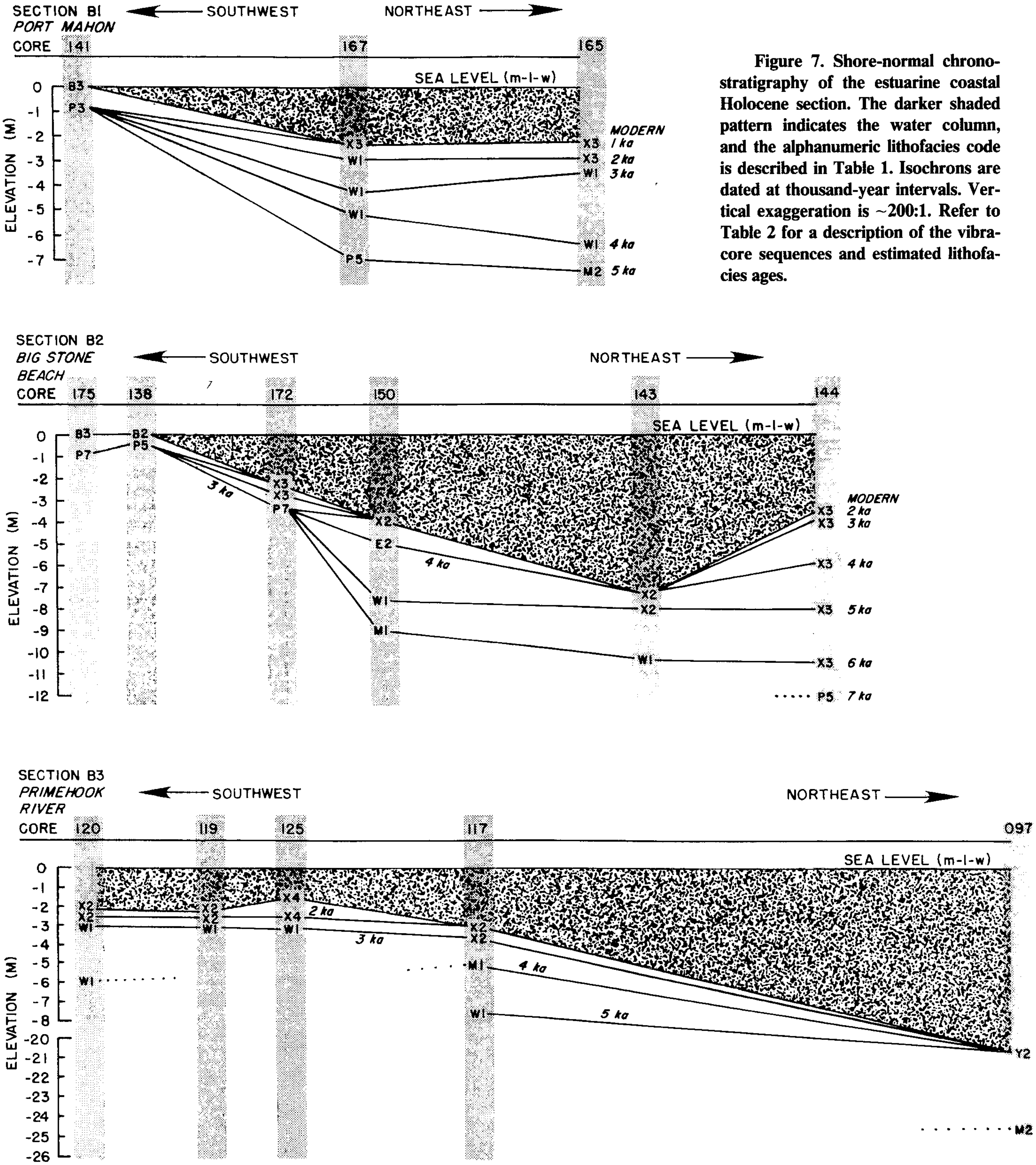


Figure 7. Shore-normal chronostratigraphy of the estuarine coastal Holocene section. The darker shaded pattern indicates the water column, and the alphanumeric lithofacies code is described in Table 1. Isochrons are dated at thousand-year intervals. Vertical exaggeration is ~200:1. Refer to Table 2 for a description of the vibracore sequences and estimated lithofacies ages.

usually composed of tidal-wetland lithofacies (Belknap and Kraft, 1977, 1981, 1985; Fletcher, 1986; Knebel and others, 1988). Pre-transgressive highlands, formerly drainage interfluves, are generally erosive on their upper surface and laterally terminate the on-lap sheets which infill the lowlands (Kraft, 1971; Belknap and Kraft, 1977, 1981, 1985; Fletcher, 1986; Knebel and others, 1988). Descriptions of the lithofacies and their estimated ages in the 27 vibracores of cross sections B1 through B4 may be found in Table 2.

Section B1 extends to the northeast from Port Mahon, across the estuarine shoreface, and consists of three vibracores. Core 141 records a

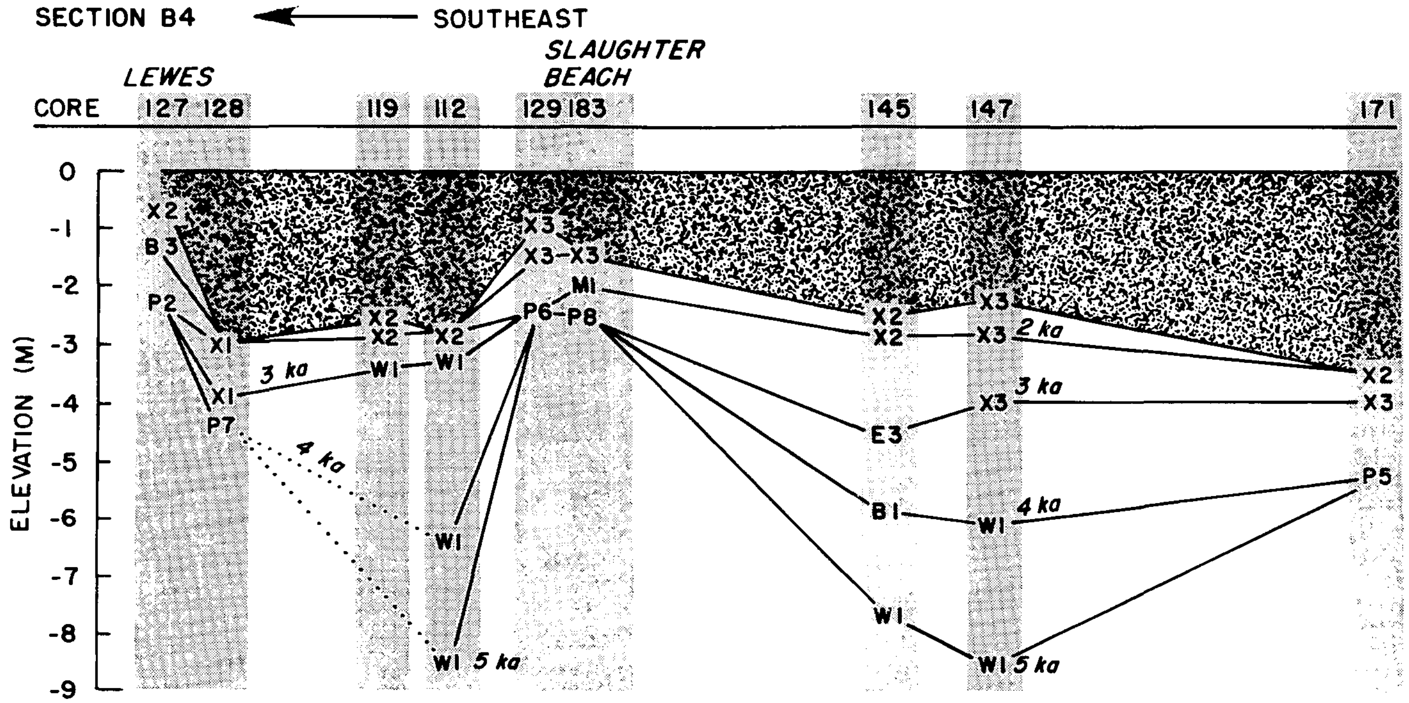


Figure 8. Shore-parallel chronostratigraphy of the estuarine coastal Holocene section. The darker shaded pattern indicates the water column, and the alphanumeric lithofacies code is described in Table 1. Isochrons are dated at thousand-year intervals. Vertical exaggeration is ~200:1. Refer to Table 2 for a description of the vibracore sequences and estimated lithofacies ages.

thin washover barrier accreted onto a pre-transgression highland. Offshore, cores 167 and 165 intersect basement lowland adjacent to the main channel of the paleodrainage system. Holocene sediments on this lowland illustrate the preservation of a thick tidal wetland sequence (underlain by basal peat) despite active erosion at the shoreface during sea-level rise. This chronostratigraphy shows the shore-normal relationship between eroding pre-transgression highland and lowland accumulations from the middle and late Holocene.

Extending northeast from the pre-transgression highland at Big Stone Beach, north of the Mispillion River, section B2 presents a more complex chronostratigraphy. Cores 175 and 138 record the basal contact at depths of less than 1 m. Landward, to the southwest, there are extensive subaerial outcrops of pre-Holocene sediments, suggesting that most of the thin Holocene section sits atop a shallowly dipping basement. Offshore, however, the pre-transgression surface is penetrated at increasingly greater depths: in core 172 at a depth of more than 3 m and in core 144 at a depth of more than 12 m. Overlying estuarine and tidal wetland sediments are preserved due to their location on a pre-transgression lowland. Clearly, basement in this transect descends across a pre-transgression interfluve into a former channel valley, probably the southern valley of Knebel and others, 1988 (see Fig. 1C). The extent of erosion is displayed in core 143, at a water depth of more than 6 m; this location is experiencing erosion to the 4 Ka isochron. The presence of a tidal shoal in core 144 provides for a thick accumulation of modern open-water lithofacies atop the pre-transgression surface. Earlier Holocene sediments appear to have been eroded during or prior to the formation of the shoal. In general, the chronostratigraphy of section B2 suggests that 3,000 to 4,000 yr of section has been eroded from the southwest estuarine shoreface, whereas the preserved section dates from the early to middle Holocene.

Section B3, along the main axis of the ancestral Primehook River, contains a well-developed Holocene sequence which locally exceeds thicknesses of 24 m. Core 97 has the deepest penetration; it samples marsh peat that has been sea-level dated to >10 Ka. The depth of nearshore erosion at cores 117 and 125 has been restricted to 1,000 to 2,000 yr. Overlying the entire section, there is a late Holocene open-water lithofacies. This transect is a counterpoint to the lesser thickness and poor preservation which characterizes Holocene accumulations atop, and along the flanks of, pre-transgression highlands.

Section B4, in Figure 8, depicts the shore-parallel chronostratigraphy of the southwestern coast. Constructed with the correlation of 15 vibracores, this section highlights a number of regionally significant chronostratigraphic relationships. Four areas exhibiting thin, poorly developed Holocene sections underlain by pre-transgression lithofacies are found at Lewes (cores 127 and 128), Slaughter Beach (cores 129 and 183 south of Cedar Creek), Big Stone Beach (cores 171, 172, and 155 north of Mispillion River), and to the southeast of Port Mahon (core 163) (see Fig. 3). The sections preserved in cores 163 and 183 are in shallow water of less than 1.5-m depth and consist of open-water lithofacies underlain by tidal-wetland—marsh lithofacies. Cores 128, 171, 172, and 155, in deeper water, contain a thin cover of estuarine sands over the pre-transgression lithofacies. This suggests that the Holocene section underlain by shallow pre-transgression lithofacies is subject to active reworking, effectively removing all Holocene lithofacies other than open-water lithofacies.

A second regionally important chronostratigraphic relationship is found in the incised stream valleys of the ancestral drainage system. During the course of transgression, these local basins were characterized by the rapid aggradation of Holocene lithofacies, as shown by cores 119, 112, 145, 147, 181, and 162, all of which fail to penetrate the pre-transgression surface despite their relatively great lengths (Table 2). Most well-developed Holocene sequences are found in these former channels. These consist of lithofacies from the middle and late Holocene which have survived the erosional processes accompanying the rise of sea level. Among

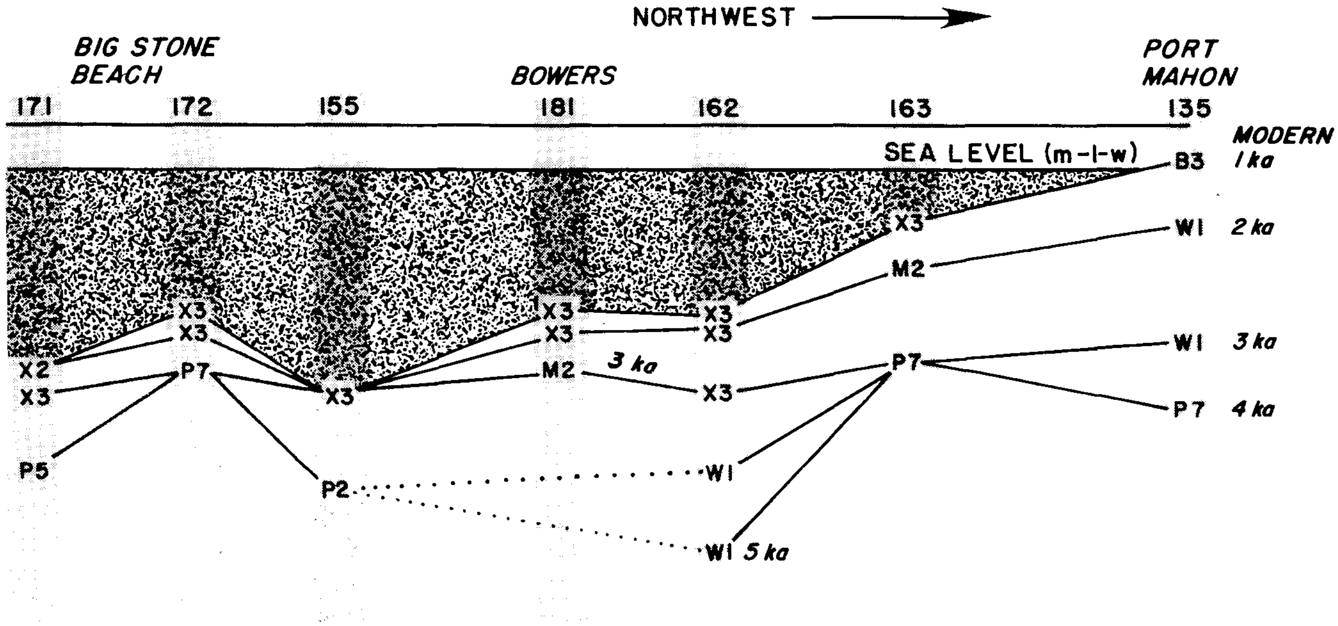


Figure 8. (*Continued*).

the salient features of Holocene valley sections are the display of thicker isochronous units, lateral termination against pre-transgression highlands, and a greater variation in the character of the preserved lithofacies.

Lithofacies Analysis

The lithostratigraphy may be readily distilled into a few primary lithofacies relationships with the technique of randomized lithofacies transition probability analysis. This method was originally used by Selley (1970) and later described in detail by Harms and others (1982). A lithofacies transition is a contact between two vertically adjacent lithofacies. The technique identifies those lithofacies which tend to occur together, and those which may be mutually exclusive or less commonly related. A nonrandom transition probability matrix is constructed, describing lithofacies contacts in the vibracore data base which have a statistical probability of being more common or less common than a random sequence of transitions. A flow diagram (Fig. 9) illustrates the important transitions and also displays the percentage of total transitions in the data base represented by each pair of lithofacies.

Three lithofacies complexes or systems are identified by probability analysis of the Delaware Bay coastal Holocene section: the accretionary cape complex, the shallow pre-Holocene erosional surface, and the pre-Holocene valley complex (Fig. 9).

The accretionary cape lithofacies complex is associated with the local progradation at Cape Henlopen. Accounting for about 14% of the transitions, this complex consists of shallow estuarine and prograding spit lithofacies. As the spit sands and gravels are about eight times more likely to be the overlying unit than are the shallow estuarine sediments, this model is consistent with the high rate of sediment influx to the Cape Henlopen area and the generally northern direction of cape progradation across formerly estuarine environments (Maurmeyer, 1974; Kraft and John, 1976; Kraft and others, 1978).

The shallow pre-Holocene erosional surface is occupied by a number of lithofacies which, together, account for nearly 24% of the transitions in the data base. In this lithofacies system, pre-Holocene basement is found at shallow depths, which, in the case of continued sea-level rise, causes the overlying Holocene section to be partially or completely eroded due to shoreface wave action (Kraft, 1971; Sheridan and others, 1974; Belknap and Kraft, 1981, 1985; Fletcher, 1986). The most likely sediment to be found atop the shallow pre-Holocene surface is either a basal Holocene gravelly sand, or an estuarine or shallow estuarine lithofacies of muddy to gravelly sands. The possibility of a previously existing tidal-wetland lithofacies having been eroded from this surface is great. Often, the open-water lithofacies are derived, in part, from the reworking of earlier Holocene or pre-Holocene sediments. The frequent occurrence of this lithofacies system and the profound control it exerts on the preservation potential of the Holocene section (Belknap and Kraft, 1981, 1985) identify it as an important component of the coastal stratigraphy.

The greatest thicknesses of Holocene sediment are found in conjunction with the pre-Holocene valley lithofacies complex. Here, 62% of the Delaware Bay coastal Holocene section is found in a sequence of tidal-wetland, barrier, and open-water lithofacies. Overlying the pre-Holocene lowland, there are basal peats, or, less commonly, there is a barrier sand which has accreted onto the basement surface and been buried by open-water deposition. The basal peat is often found with other types of tidal-wetland lithofacies, which in turn may be overlain by washover sands. The most common sequence, however, includes a tidal-wetland section of basal salt-marsh peat overlain by an accumulation of organic mud, which in turn has been overwashed by barrier migration and is currently exposed in the shallow estuary as an eroding subtidal mudflat (Weil, 1976, 1977; Maley, 1981; Marx, 1981; Fletcher, 1986). These lithofacies transitions denote the partial preservation of the Holocene coastal section, as the exposed tidal-wetland muds are typically more than 1,000 yr old and were previously overlain by back-barrier and barrier sediments that have subsequently migrated landward or been removed by transgression-related erosion.

Aggradation and Erosion of the Coastal Section

The texture and composition of the tidal-wetland lithofaces (Khalequzzaman, 1989; Whallon, 1989) and the recognition that processes controlling sediment distribution across the salt marsh originate from the adjacent estuary (Postma, 1967; Elliott, 1972; Richter, 1974; Allen, 1977;

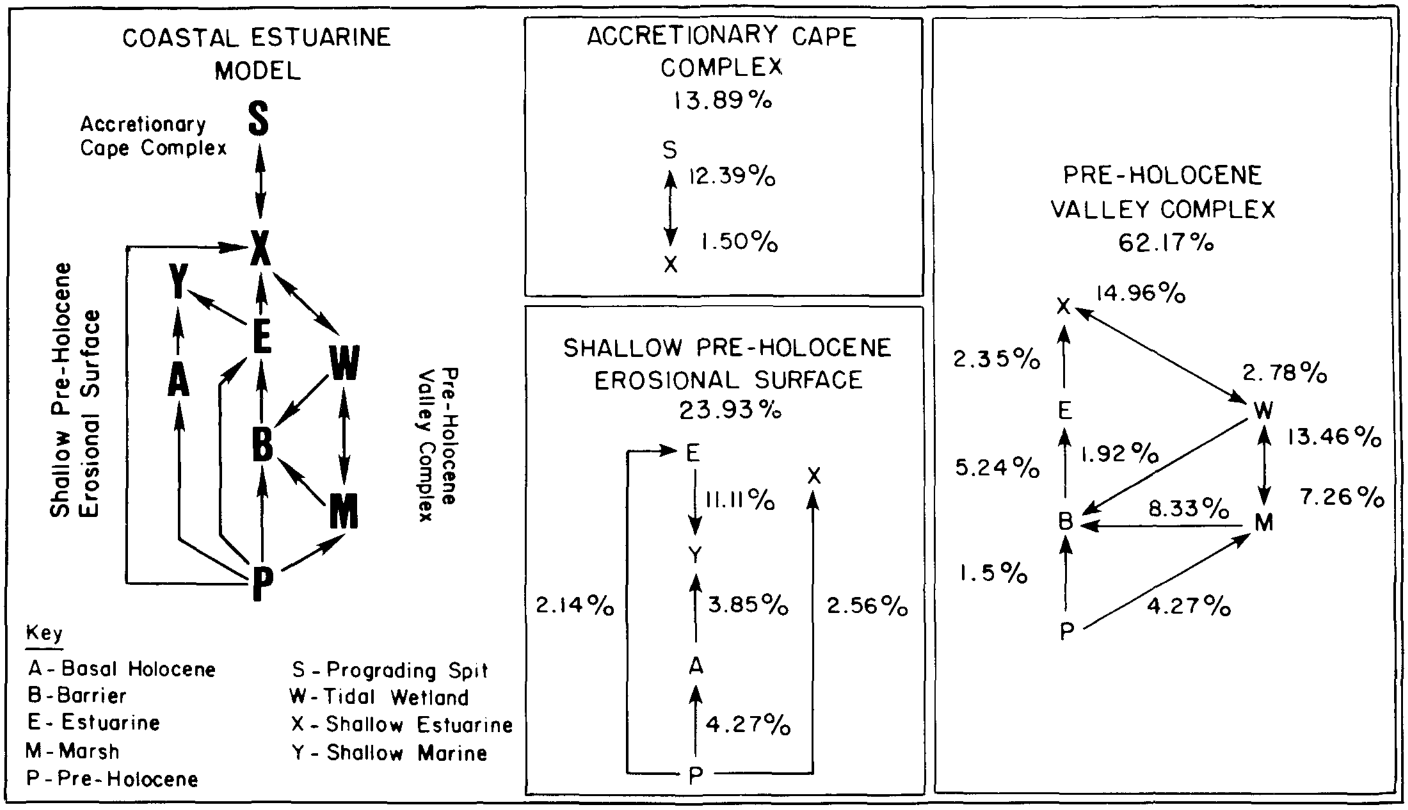


Figure 9. Flow diagram showing results of lithofacies transition analysis of the Delaware Bay coastal section.

Stumpf, 1981; Frey and Basan, 1985) suggest that suspended sediment from the estuarine turbidity maximum is being actively deposited on the marsh surface. The organic and inorganic silt and clay of the tidal-wetland lithofacies has originated, at least in part, from seston concentrations in an adjacent tributary tidal stream or from the axis of the main Delaware River channel. The combined influence of the tributary and central channel turbidity maxima creates a regional depocenter which migrates with rising sea level at the head of the estuarine system. The history of coastal stratigraphic aggradation is therefore tied to the history of this turbidity maximum depocenter (Fig. 10), which, during the Holocene, has moved northwest across the estuarine basin (Kraft, 1971, 1979; Kraft and others, 1973, 1979; Richter, 1974; Weil, 1976, 1977; Maley, 1981; Marx, 1981; Fletcher, 1986; Knebel and others, 1988).

The Holocene coastal sequence began to aggrade as a result of the tidal influx of sediments from this proximal turbidity maximum depocenter (Weil, 1976, 1977; Fletcher, 1986, 1988a; Fletcher and Pizzuto, 1986). The sediments of the pre-Holocene valley lithofacies complex compose the largest part of the coastal strata. These lithofacies were initially deposited coincident with the flooding of the estuarine basin and the influence of the proximal depocenter. Depocenter activity was both local in nature, as in the case of salt-wedge intrusion along a tributary valley, and more regionally extensive, as along the main axis of the estuary. Fine-grained sediments comprising the depocenter accumulated both proximal to the turbidity maximum and distally, following extended tidal transport. Weil (1976, 1977) described these deposits as part of an estuarine delta, developing in response to sediment influx to the depocenter from upland sources, followed by fine-sediment distribution and deposition by tidal circulation along the estuarine margin. As sea level rose through the Holocene, the first stage of inundation was characterized by aggradation of tidal-wetland lithofacies (Fig. 11), forming the lower part of the coastal section. This has been named the "constructive deltaic phase" by Weil (1976). As the estuary became wider and deeper, the aggraded tidal-wetland lithofacies strata were truncated and eroded in conjunction with the accumulation and landward migration of barrier-spit lithofacies. This is known as the "destructive deltaic phase" (Weil, 1976). Evidence of the erosion is found in the form of extensive subtidal flats ringing the margin of the lower estuary (Weil, 1976, 1977; Maley, 1981; Marx, 1981; Fletcher, 1986; Knebel and others, 1988). Holocene aggradation may continue with the deposition of open-water lithofacies atop the partially or wholly eroded coastal section (Sheridan and others, 1974; Weil and others, 1975; Weil, 1976, 1977; Fletcher, 1986; Knebel and others, 1988; Knebel, 1989).

Radiocarbon dates of organic tidal-wetland sediments (Kraft, 1976; Maley, 1981; Marx, 1981; Belknap and Kraft, 1985) and the chronostratigraphy derived from the relative sea-level age estimation of lithofacies in the vibracore data base allow us to present a history of the migrating depocenter (Fig. 10). Prior to about 10 Ka, the depocenter was located on the inner continental shelf southeast of the present mouth of Delaware Bay (Figs. 10 and 11). The modern estuary basin was then part of the regional fluvial system of the ancestral Delaware River, and the estuary at that time was located to the east, on the modern continental shelf. Between 10 Ka to 8 Ka, rising sea level had moved the depocenter along the main axis of the ancestral channel to approximately the location of the modern bay mouth. The coast between Cape Henlopen and the Mispillion River entered the constructive deltaic phase of sedimentation at this time (Fig. 11).

The middle Holocene record shows the movement of the depocenter to the northwest and the continued aggradation of the coastal section in the Broadkill, Primehook, Cedar Creek, Mispillion, and Murderkill valleys. Between 5 Ka and 3 Ka, the easternmost portion of the southwest estuarine coast experienced the erosive effects of the wind-wave climate resulting from the greater fetch of the broadened estuary. Following 3 Ka, the depocenter had migrated beyond the modern estuarine basin. This later phase of sedimentation, the destructive deltaic phase, is characterized by the end of proximal aggradation of the coastal section with the onset of barrier migration and subsequent tidal-wetland exposure to erosion. Modern sedimentation occurs landward of the barriers with continued distal aggradation of the tidal-wetland lithofacies, and in the open estuary with the accumulation of open-water lithofacies. Both Kraft and others (1989) and Church and others (1987) have determined that modern tidal-wetland sedimentation rates average about 4 to 5 mm/yr, a factor of three or more higher than the long-term average of the past 1,000 yr. The reason for the high modern sedimentation rates is problematic and may be due to several factors, including basin subsidence, compaction, increased sea-level rise, increased storminess, and increased sediment availability.

COASTAL EVOLUTION

Tide Dominated to Wave Dominated

The observation that coastlines exhibit distinct morphologies in response to the dominance of either wave or tide energies was first reported by Price (1955). This approach to understanding the occurrence of specific coastal morphotypes through an understanding of formative processes was continued with the work of Davies (1964) and Hayes (1975, 1979). In the case of deltaic systems, coastal morphogenesis is complicated by influence of a third category of incident energies, the fluvial system (Galloway, 1975; Wright, 1978).

Tide-dominated and wave-dominated coasts display variations in inlet frequency, ebb delta development, barrier morphology, and longshore features which are reflections of processes that transport sediment (Davis and Hayes, 1984). The complex of morphodynamics in a coastal setting produces a specific stratigraphy which allows interpretation of the paleocoastline (Clifton and Dingler, 1984; Dupré, 1984).

Along the southwest coast of Delaware Bay, the constructive deltaic phase of sedimentation occurred in conjuction with tide-dominated distribution of suspended sediments in the turbidity maximum depocenter. The destructive deltaic phase is characterized, in part, by high-energy erosion of shoreline and nearshore deposits by predominantly wave-related processes. The observed relationships of dominant process, phase of sedimentation, and depositional environments across the estuary can be expanded to describe the major areal environmental variations and response of processes occurring throughout the estuary during the transgression (Fig. 12).

Lithofacies accumulations on the estuary coast reflect the changing influence of tide versus wave-related processes. Above the basal pretransgression lithofacies, strata consist largely of thick sections of tidal-wetland muds (Richter, 1974; Kraft and John, 1976; Maley, 1981; Marx, 1981; Fletcher, 1986). These sediments were deposited, for the most part, during the early stages of inundation when tidal distribution was the primary mode of sediment dispersal. At that time, the coast was characterized by muddy shorelines and an absence of significant accumulations of sediment coarser than fine sand. Due to limited fetch in the estuary, wind-wave activity was minimal.

During later stages of inundation, the greater fetch of the open estuary produced an increase in wind-wave energy. Coarse-grained material, the barrier-spit lithofacies, accumulated in the form of migrating washover barriers and eroding headland beaches. Tidal currents were still an influential factor in the distribution and reworking of open-estuarine sediments, but the morphology of the coast and character of the accumulations were profoundly changed by the increasing influence of the higher energy wind-wave regime.

The change from one mode of coastal morphodynamics to another was gradational (Fletcher and Kraft, 1986a). The relative influence of tide-to-wind–dominated processes was part of a continuum produced by rising sea level (Fletcher and Kraft, 1986a). The two phases of estuarine coastal sedimentation which resulted in the final configuration of Holocene deposits may be characterized by this process continuum. The constructive deltaic phase of sedimentation occurred in conjunction with proximal tide-dominated distribution of suspended sediment. The destructive deltaic phase is characterized by distal sedimentation and the high-energy erosion of coastal deposits by predominantly wave-related processes.

Implications

During the Holocene transgression along the estuary coast, the phase relationship between the aggradation and erosion of the Holocene section

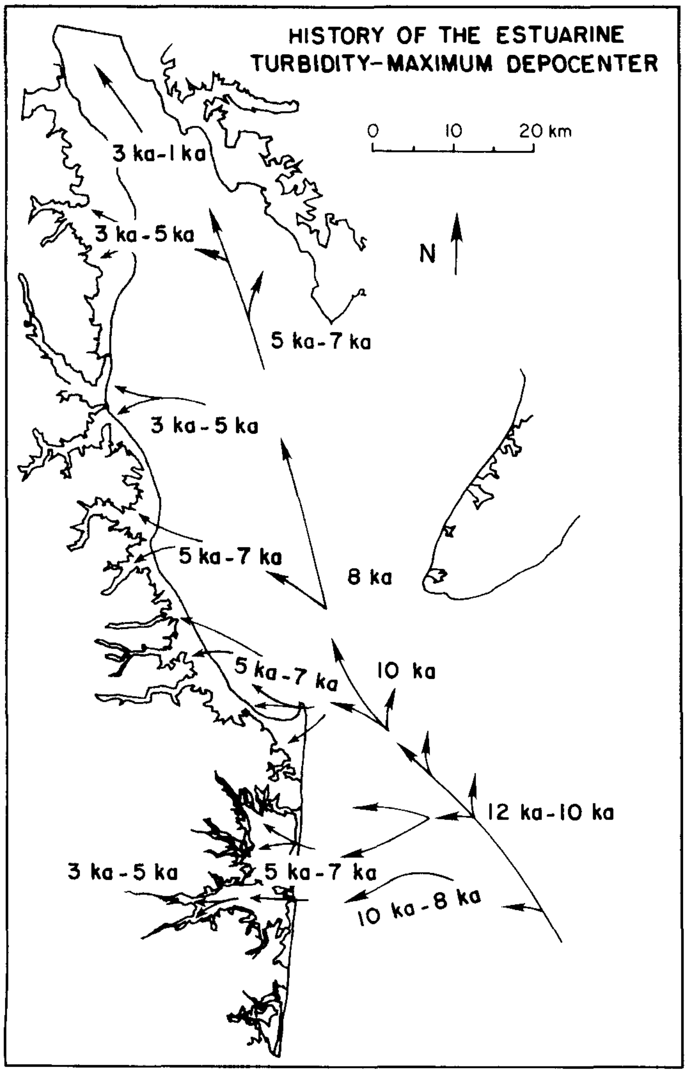


Figure 10. History of the migrating, estuarine turbidity-maximum depocenter along the axes of the ancestral Delaware River and major tributary systems.

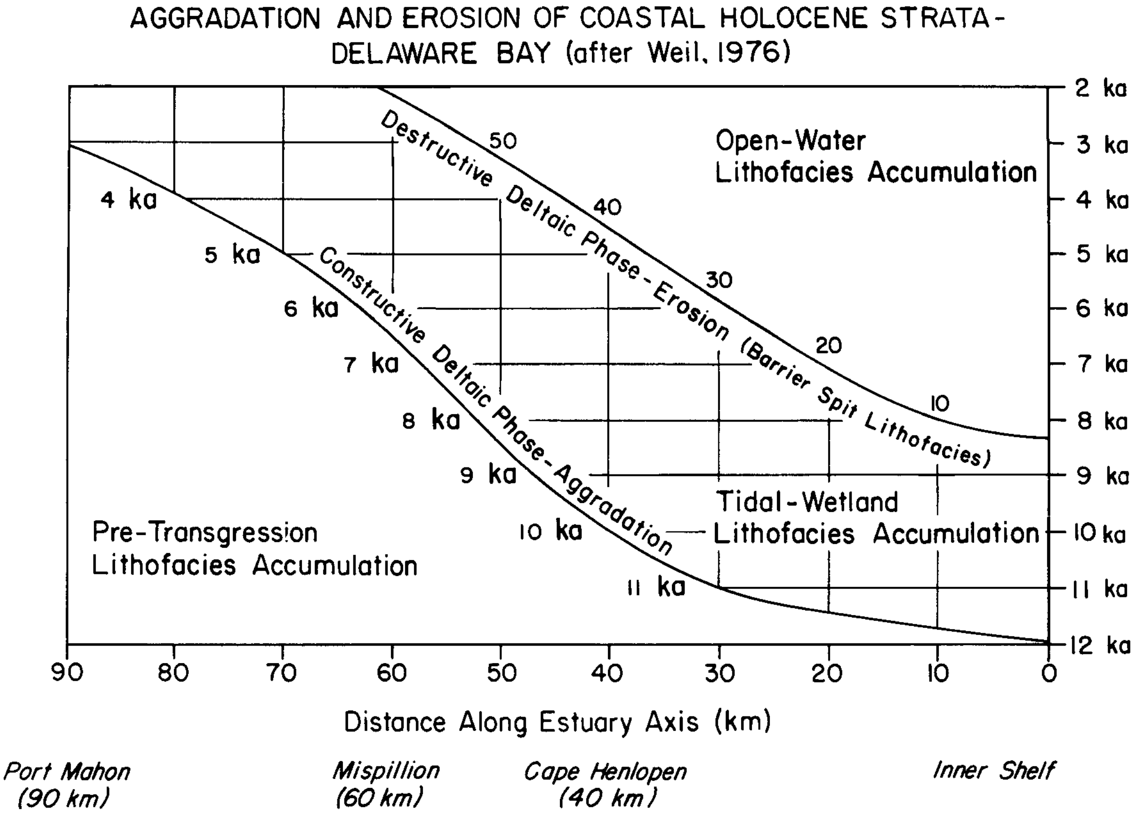


Figure 11. Aggradation and erosion of coastal Holocene strata during the constructive and destructive phases of estuarine delta sedimentation. The hachured volume represents the residence time and preservation potential of strata along the Delaware Bay southwest coast. Prior to inundation by rising sea level, pre-transgression lithofacies accumulate as basal Holocene and pre-Holocene deposits. The aggradation of coastal Holocene strata begins with the start of the constructive deltaic phase, characterized by proximal deposition of turbidity-maximum sediments. Tidal-wetland lithofacies accumulate until wind-wave energy at the shoreface causes the landward migration of barrier-spit lithofacies. This initiates the destructive deltaic phase of sedimentation, characterized by erosion of the back-barrier tidal-wetland deposits by migrating washover barriers and subsequent exposure on the shoreface as a subtidal mud flat. Open-water lithofacies may then accumulate on this surface above the truncated coastal strata.

varied in time and space (Fig. 11). Several factors influenced this relationship. These include (1) rate of sediment influx to a particular site, (2) energy incident to the shore, (3) exposure to prevalent low-pressure weather systems, (4) activity of tidal currents, (5) rate of sea-level rise, and (6) configuration of the pre-transgression surface (Belknap and Kraft, 1977, 1981, 1985).

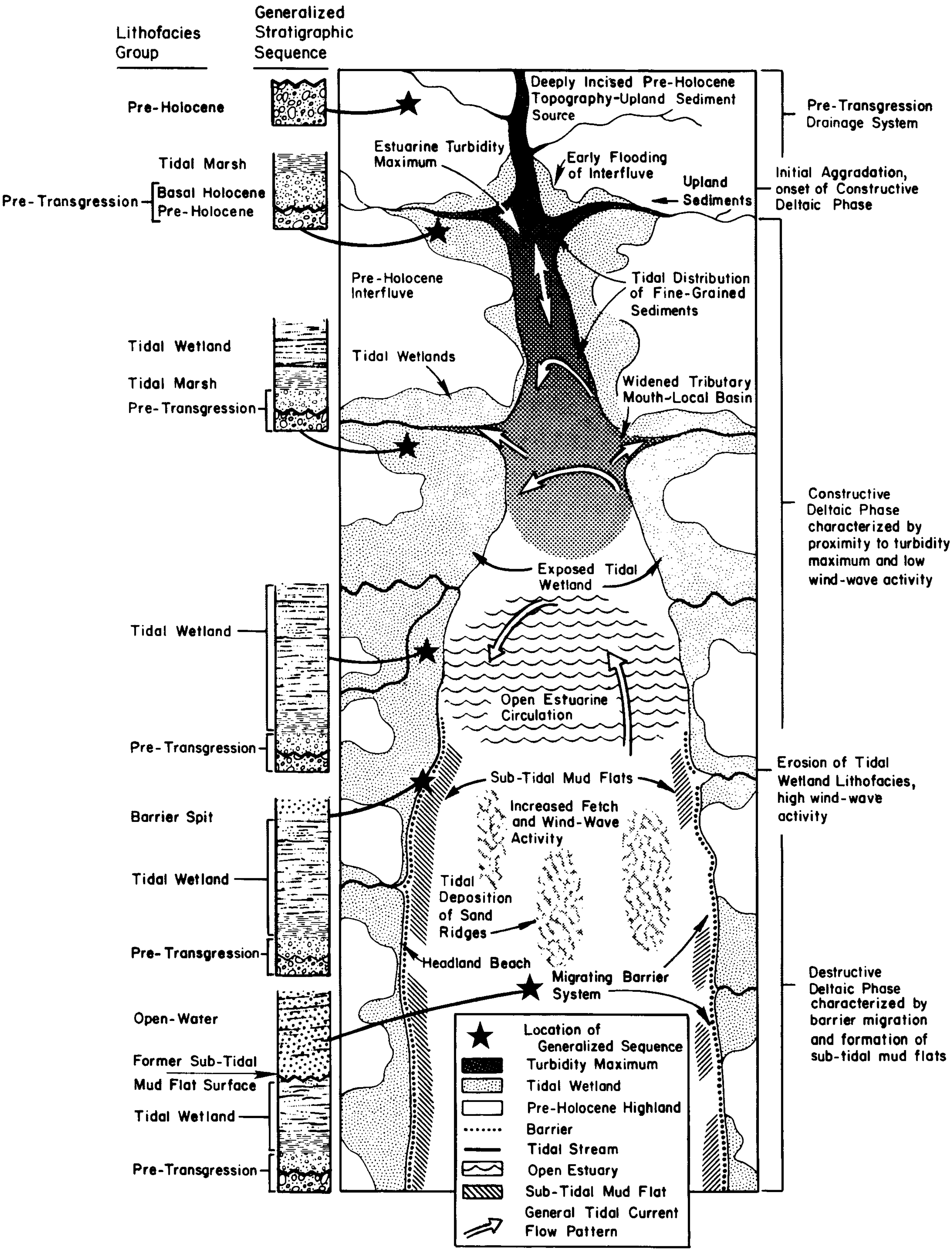
The history of aggradation and erosion of coastal strata is controlled by the constructive and destructive phases of deltaic sedimentation (Fig. 11). The duration of Holocene aggradation is the temporal span between the first influence of the turbidity maximum depocenter at a locality, and the onset of wind-wave erosion. The erosion extends to a depth equal to storm wave base followed by deposition of a variable thickness of open-water lithofacies accumulating in response to tidal current activity along the estuarine floor (Fig. 12).

In Figure 11, the time between the initiation of aggradation and the beginning of erosion is the period over which the section develops; it may be thought of as the residence time of the section. The slope change in the line of aggradation is probably related to changes in the rate of sea-level rise. The line of erosion, however, is probably more a result of developing wind-wave energy and antecedent topography than the actual rate of sea-level rise. The time difference between the two averages ~4,000 yr. Erosion of the section does not result in total loss of Holocene materials. In those cases, where the pre-Holocene valley complex extends deep enough to survive storm reworking, several thousand years of early to late Holocene strata have survived (Belknap and Kraft, 1981; Maley, 1981; Marx, 1981; Fletcher, 1986; Knebel and others, 1988). Shoreface erosion does not remove tidal-flat and sand-ridge deposits of the deep bay.

Summary of Evolutionary History

During the Holocene transgression, a regional depocenter, related to tributary and main-channel turbidity maxima, migrated northwest across the estuarine basin at a rate determined by rates of sea-level rise and along a path determined by the topography of the pre-transgression surface. By

Figure 12. Coastal evolutionary history, stratigraphic development, sediment source and dispersal systems, and inundation patterns in the Delaware Bay estuary over the course of the Holocene transgression.



Lithofacies Group
Generalized Stratigraphic Sequence
Pre-Holocene
Tidal Marsh
Pre-Transgression
Basal Holocene
Pre-Holocene
Tidal Wetland
Tidal Marsh
Pre-Transgression
Tidal Wetland
Pre-Transgression
Barrier Spit
Tidal Wetland
Pre-Transgression
Open-Water
Former Sub-Tidal Mud Flat Surface
Tidal Wetland
Pre-Transgression
Deeply Incised Pre-Holocene Topography-Upland Sediment Source
Estuarine Turbidity Maximum
Early Flooding of Interfluve
Upland Sediments
Pre-Holocene Interfluve
Tidal Distribution of Fine-Grained Sediments
Tidal Wetlands
Widened Tributary Mouth-Local Basin
Exposed Tidal Wetland
Open Estuarine Circulation
Sub-Tidal Mud Flats
Increased Fetch and Wind-Wave Activity
Tidal Deposition of Sand Ridges
Headland Beach
Migrating Barrier System
Location of Generalized Sequence
Turbidity Maximum
Tidal Wetland
Pre-Holocene Highland
Barrier
Tidal Stream
Open Estuary
Sub-Tidal Mud Flat
General Tidal Current Flow Pattern
Pre-Transgression Drainage System
Initial Aggradation, onset of Constructive Deltaic Phase
Constructive Deltaic Phase characterized by proximity to turbidity maximum and low wind-wave activity
Erosion of Tidal Wetland Lithofacies, high wind-wave activity
Destructive Deltaic Phase characterized by barrier migration and formation of sub-tidal mud flats

12 Ka, this depocenter was located southeast of the modern Bay mouth. By 10 Ka, the depocenter had relocated to approximately the position of the modern Bay mouth, resulting in sediment dispersal along the channel axes of the ancestral Delaware River drainage system in the area. On the southwest coast, exposed tidal-wetland muds occupied widened tributary mouths and extended peripherally around flooded interfluves during the period 7 Ka to 5 Ka. By the middle to late Holocene, the southeastern portion of the estuary had widened, and heightened wind-wave activity modified the southwest shore and caused erosion of existing deposits (Figs. 10, 11, and 12).

The influence of changing sea level and sediment availability caused the estuarine coast to constantly adjust to new conditions, describing a continuum between the ancestral fluvial system and the completely inundated estuarine floor.

Initially, a rise in sea level resulted in flooding of the land surface and tidal deposition of organic fine-grained sediments. Local tidal fresh-water and fluvial wetlands soon gave way to tidal brackish wetlands which extended across the pre-transgression surface, creating a mud coast occupied by salt-tolerant vegetation in a low-energy, open-water environment. This muddy, vegetated coast was originally marked by gravelly highlands, but high rates of fine-grained sediment influx expanded local tidal wetland deposits until they coalesced laterally, creating long expanses of shoreline consisting exclusively of tidal wetlands.

With continued sea-level rise, the estuary broadened, and the increasing fetch resulted in a high level of wind-wave energy at the shore. Consequently, wave action against the exposed tidal wetlands and highlands led to the development of washover barriers and headland beaches. As these transgressive coarse-grained deposits moved landward, subtidal mud flats, composed of tidal wetland lithofacies, were exposed along the estuarine margin. Eventually, these muddy surfaces, the eroded remnants of the coastal stratigraphic section, were covered by estuarine deposits of the open-water lithofacies group.

The coastal Holocene section developed on the estuarine margin during an initial period of aggradation associated with proximity of the turbidity maximum, the constructive deltaic phase. Aggradation, lasting for ~4,000 yr, consisted of three interrelated lithofacies complexes: the pre-Holocene valley complex, the accretionary cape complex, and the shallow pre-Holocene erosional surface. With continued sea-level rise and broadening of the expanse of the estuary, heightened wind-wave energy combined with distal sedimentation patterns to initiate the destructive deltaic phase. The dominant features of this period of coastal evolution are the landward migration of washover barriers and the simultaneous truncation of the tidal wetland accumulations. Deposition of open-water lithofacies on the muddy shoreface marks the end of coastal evolution and the beginning of a regime of estuarine sedimentation.

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