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Coastal Processes and Offshore Geology

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ABSTRACT

The modern coastal geology of Virginia results from the interactions of modern processes, primarily waves, tidal currents and sea-level rise, with the antecedent geology. The ancient and major rivers draining the Piedmont and interior highlands of eastern North America carried sediments that were deposited in various areas across the physiographic continuum of the coastal plain and continental shelf as sea level fluctuated in response to global climate changes. The scarps that formed by shoreline erosion during highstands of sea level and the very low-gradient intervening flats are the proximal underpinning of the contemporary coastal zone. The ocean shoreline of Virginia comprises parts of two major coastal compartments: one spanning the distance between Delaware and Chesapeake Bays; the other running from Cape Henry, the southern side of the mouth of the Chesapeake, to Cape Lookout, North Carolina. The location within the broader coastal compartment and the local interplay of the processes with the geography determine the development of the shoreline within each segment of the shore.

The gross characterization of Virginia's coast as the Delmarva Peninsula, the Bay Mouth, and Southeastern Virginia sections insufficiently describes the variation. The two major subaerial compartments can be further segmented. The Delmarva Peninsula embraces, from north to south, the Cape Henlopen spit complex, the eroding headlands of Bethany and Rehoboth, the long Fenwick-Assateague barrier island terminating in a potentially developing cape adjacent to Chincoteague Island, the Virginia Barrier Islands, and the distal Fishermans Island at Cape Charles. The Virginia Barrier Islands can be partitioned into a wave-dominated, severely eroding northern segment, a central transitional segment, and a southern segment with greater tidal influence. The Chesapeake Bay Mouth is a complex region of shoals and channels responding to wave energy and reversing tidal currents flowing into Chesapeake Bay. The Southeastern Virginia compartment mimics Delmarva with the northern spit of Cape Henry, the Virginia Beach headland, and the long barrier-island complex that continues to Cape Lookout, where the shoreline turns sharply west toward the mainland.

INTRODUCTION

The coastal plain and continental shelf of Virginia are contiguous, but discrete, physiographic provinces delimited by the present elevation of sea level. However, on geologic time scales of thousands to tens of thousands, and even millions, of years, the *coastal zone* – the boundary between the coastal plain and shelf – is dynamic and migrates freely hundreds of kilometers landward and seaward across the continental margin. The modern coastal zone occupies nearly the same position as several previous interglacial highstands of sea level that have recurred at approximately 100,000-year (abbreviated 100 ky, for "kilo year") intervals since the Middle Pleistocene (about the last 750 ky).

The Virginia coastal zone encompasses the outer coastal plain and the barrier islands and mainland shoreline fronting the Atlantic Ocean, as well as the back-barrier lagoons or coastal bays, the inner shelf, and the mouth of Chesapeake Bay (Figure 1). A broader definition may include the lower reaches of the estuarine tributaries to the southern part of Chesapeake Bay, such as the James, York, and Rappahannock Rivers.

Within the coastal zone, the geomorphology (landforms), both emergent and submerged, are the

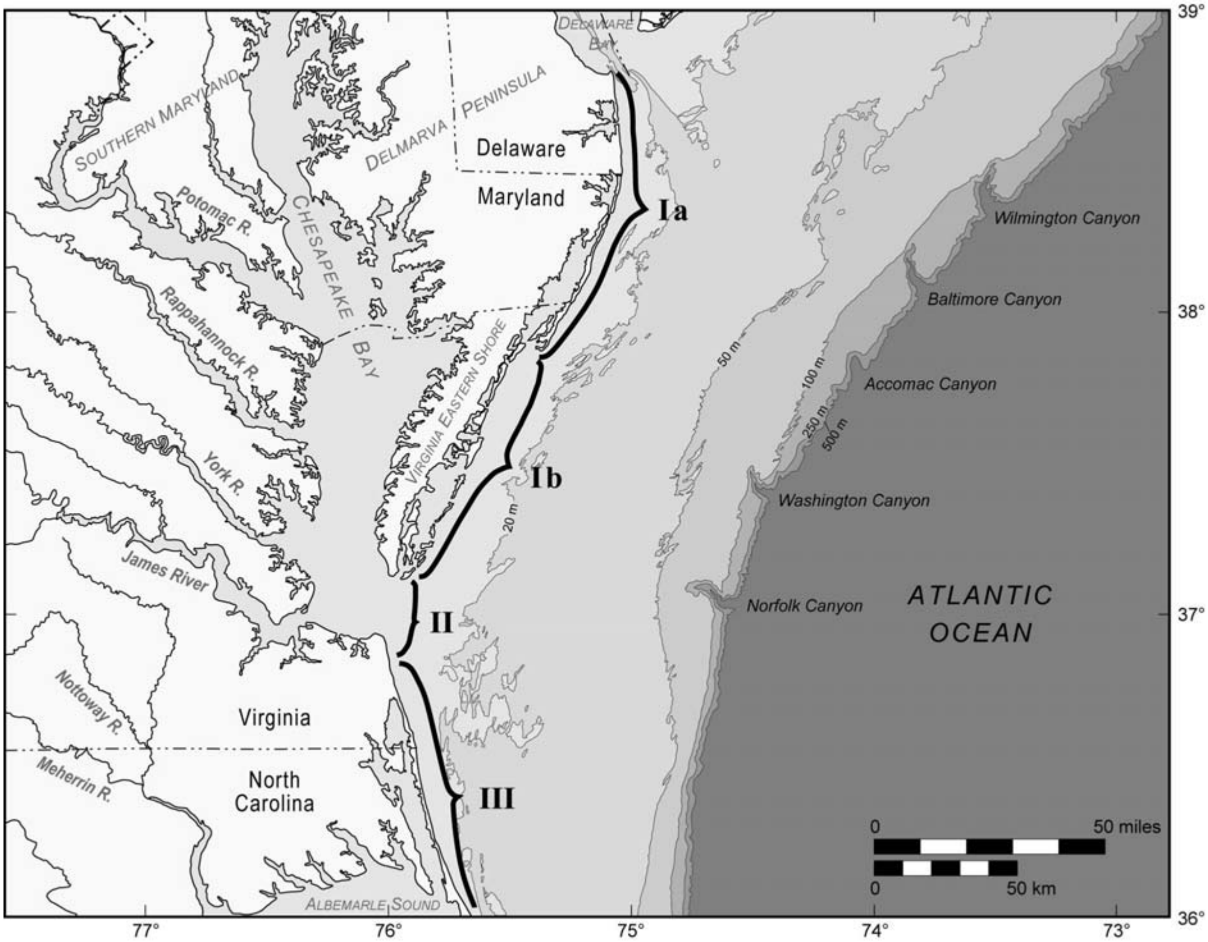


Figure 1: The Virginia coastal zone can be divided by geologic character into three broad geographic areas: the Atlantic coast of the Delmarva Peninsula, the mouth of Chesapeake Bay, and the mainland shore of southeastern Virginia continuing into northeastern North Carolina. Each of these main sections can be subdivided further into compartments defined primarily by the effects of wave and tide energy. The Southern Delmarva section includes three smaller segments: the arc of severe erosion from Chincoteague Inlet to Wachapreague Inlet, the central section of Parramore, Hog, and Cobb Islands, and the arcuate reach of Wreck through Smith Islands which follows the inner rim of the Chesapeake Bay Impact Crater.

result of a geologic heritage modified by modern processes driven by the energy of waves and tides. High wave energy from the Atlantic Ocean shapes the mainland shore of southeastern Virginia and the long, narrow barrier islands of the central Delmarva (**Del**aware-**Mar**yland-**V**irgini**a**) Peninsula. Mixed wave and tide energy create the segmented barrier islands of the Virginia Eastern Shore (the southern extension of the Delmarva Peninsula) that are separated by deep tidal inlets and accompanying tidal deltas. Strong flood and ebb currents move water and sand through the mouth of Chesapeake Bay, creating alternating deep channels and shallow shoals between the Virginia Capes – Cape Charles on the southern tip of the Virginia Eastern Shore and Cape Henry on the northern edge of Virginia Beach. Throughout the coastal zone, gradients of energy control the net transport of sediment, producing in some areas smooth, gradual transitions in coastal geomorphology, and in other areas, abrupt changes in character.

The Virginia coast cannot be viewed as an isolated entity, arbitrarily bounded by state lines, but as part of a larger regional system with a shared geologic history. The Virginia coastal zone can be divided by geologic character into three broad geographic areas: the Atlantic coast of the Delmarva Peninsula, the mouth of Chesapeake Bay, and the mainland shore of southeastern Virginia continuing into northeastern North Carolina (Figure 1). Each of these main sections can be subdivided further into compartments defined primarily by the effects of wave and tide energy. One fundamental property of the Virginia coastal zone is that it is composed entirely of unconsolidated sediments, such as sand and silt, with no exposures of bedrock or hard, consolidated sediments. Consequently, sedimentary processes – erosion, transport, and deposition – are active on timescales of minutes to millennia and are constantly reshaping the coast.

Coastal Compartments

Fisher (1967) recognized that four large sections of the Mid-Atlantic coast, each bounded by the mouths of major estuaries, all have similar geomorphic segmentation. These sections are the coastlines of Long Island, New Jersey, Delmarva, and Virginia-North Carolina, bounded respectively by the mouths of Long Island Sound, Hudson River Estuary, Delaware Bay, and Chesapeake Bay; the southern boundary of the Virginia-North Carolina section is Cape Lookout, where the coastline turns abruptly west. Swift (1969), Belknap and Kraft (1985), and McBride (1999) used this framework to compare the coasts of New Jersey and the Delmarva Peninsula, and to explain the compartments in terms of geologic setting and sediment transport processes.

Oertel and Kraft (1994) in summarizing the coastal geology of New Jersey and Delmarva, refined Fisher's (1967) model such that each of the large sections has four segments: (1) cuspate spit, (2) eroding headland, (3) barrier spits and long linear barrier islands (wave-dominated barrier islands), and (4) short barrier islands with tidal inlets (mixed-energy barrier islands). Gutierrez *et al.* (in press) depict each of the segments within the four major compartments of the Mid-Atlantic region.

Along the entire Mid-Atlantic coast, net longshore transport of sand in the *shoreface* – the inner shelf affected by moderate waves, generally to a water depth of about 20 m – is to the south or southwest. The net southerly transport is set up by seasonal variations in the regional weather systems (Belknap and Kraft, 1985). Gentle breezes from the south and southeast predominate during the summer, but stronger winds blow from the west and northwest in winter. However, coastal storms of the late fall through early spring are the dominant force driving sand transport. Storm winds blowing from the northeast quadrant produce extremely strong south-flowing longshore currents that move more sand in a few days than is moved by less vigorous waves and currents throughout the rest of the year. Wave transformation and refraction over shallow, local features, such as shoals and tidal deltas, result in local reversals in the longshore transport.

The regional trend of southerly shore-parallel sediment transport is altered by the mouths of the major estuaries, where the gradient of wave energy goes from high along the Atlantic coast to low within the estuary. Additionally, shore-perpendicular tidal currents are focused and strengthened by constriction at the bay mouths. As a result, net sediment transport is from the ocean into the estuary (Meade, 1969, 1972). Bathymetric effects, reversing tidal flow, and interactions between waves and tides result in sediment convergence along the northern flank of the bay mouth, producing large shoal systems such as Middle Ground and Nine Foot Shoals at the mouth of Chesapeake Bay (Ludwick, 1970, 1975; Swift, 1975; Oertel and Overman, 2004). Net transport into the bay also produces a reversal of longshore transport along the ocean shoreline south of the bay, with northward flow from a nodal zone north of False Cape (Everts *et al.*, 1983). This sand moving north and then west along the southern flank of the Chesapeake Bay mouth has created the cuspate foreland of Cape Henry and Willoughby Spit prograding into the mouth of the James River estuary.

A Hierarchy of Processes in Time and Space

The present configuration of the Virginia coast has been created by processes operating since the rifting of the supercontinent Pangaea to form the North Atlantic Ocean beginning at about 225 Ma ("mega annum" or million years ago). These processes have provided sediment to the coastal zone, either as an influx of new siliciclastic sediment from the Piedmont and Appalachian Mountains or by erosion and reworking of existing sediments on the continental margin, and have produced *accommodation space* (basins) into which the sediment can be deposited. In approximate order, from large spatial and long temporal scales to processes operating today, these primary factors are the:

- structural setting of the continental margin
- foundation of Tertiary marine and marginal-marine sediments
- long-term vertical tectonic movement
- basin created by the Chesapeake Bay Impact Crater
- positions of the major rivers and their broader valleys through time
- major changes in sea level during the Pliocene and Pleistocene
- last interglacial-glacial cycle of sea level
- regional post-glacial isostatic adjustment
- continuing, and possibly accelerating, Holocene sea-level rise

The Virginia coastal plain and shelf overlie the Salisbury Embayment, a sedimentary basin located between the South New Jersey Arch and Norfolk Arch (Ward and Strickland, 1985). These structural highs and the intervening basin directly influenced Tertiary and Quaternary stratigraphic development along the Mid-Atlantic coast. Deposition in the Salisbury Embayment has produced a seaward-thickening wedge of sediments beneath the coastal plain, shelf, and continental slope. The older, deeper sediments in this sequence reflect continental depositional environments (fluvial and deltaic) that dominated the early stages of the post-rift history of the margin. These were followed by inundation by the ocean and deposition of marine and marginal-marine sediments since the Late Cretaceous.

Through the past ~160 My (million years), the persistent transfer of mass from the eroding Appalachians and Piedmont to the continental shelf and slope has produced a slow net uplift of the coastal plain landward of a hinge line near and approximately parallel to the present shoreline (Owens and Gohn, 1985; Poag, 1985; Ward and Strickland, 1985; Ator *et al.*, 2005). This vertical movement was enhanced by the general long-term subsidence of the continental margin as the bedrock cooled after rifting, and the weak transition between continental and oceanic crust sank beneath the added mass of deposited sediments (Gardner, 1989). Even as the margin was subsiding, global sea level generally dropped through the Tertiary as plate tectonics moved Antarctica slowly and steadily toward the South Pole, global climate cooled, and large ice sheets developed near both poles. The net result has been flooding of the margin by the ocean for much of the time from the late Cretaceous into the Pliocene, from approximately 100 Ma to 3 Ma, punctuated by major drops in global sea level associated with periods of growth of the Antarctic Ice Sheet, starting about 28 Ma in the early Oligocene. There has been a long-term, secular drop in sea level from a peak during the middle Pliocene at 3.5-3.2 Ma, which may have been 30-40 m higher than present (Krantz, 1991), that flooded onto the Piedmont; the shoreline of this highstand approximately follows the Interstate 95 corridor along the western edge of the coastal plain. For the past 2.5 My, with the onset of large-scale glaciation in the Northern Hemisphere, the coastal plain and shelf have been sculpted by large-amplitude, 120 to 150 m, rises and falls of sea level.

The major rivers that cross the Piedmont as they flow to the Atlantic have occupied the same valleys for millions of years. The broad valleys are separated from each other by interfluves which stand in positive relief. On the inner coastal plain, these high uplands are commonly capped by fluvial

Miocene and Pliocene sands and gravels deposited by the ancestral rivers draining the Piedmont and Appalachians. Pazzaglia (1993) depicts the James, York, Rappahannock, Potomac, Patuxent, Susquehanna, and Delaware Rivers in their northwest-southeast valleys crossing the Piedmont at least 25 My ago in the late Oligocene. Through most of the late Pliocene and Pleistocene, the rivers cut through the pre-existing coastal plain sediments and transported the eroded sediments toward the sea. Colman (1983) documented terraces along the Rappahannock River between Fredericksburg and Port Royal that are "no younger than the early Quaternary" (that is, probably older than 1 My).

The prism of sedimentary deposits that form the continental shelf of the Mid-Atlantic coast of North America began to develop during the Cretaceous and continued to grow through the Tertiary and into the Quaternary. Uchupi (1970) characterized the geological history of the continental shelf as the up-building and out-building of sediments on the continental slope over Triassic and Jurassic strata that were subsiding. The upper stratigraphic section of the shelf bears witness to the substantial variations in sea level that accompanied the major glacial episodes that began in the Pliocene. In interpreting the geology of the shelf, it is important to remember that the shelf and the coastal plain of the Mid-Atlantic are a continuum wherein demarcation between the two provinces moves with sea level. During low stages of sea level associated with glacial maxima, most of the continental shelf was emergent and, hence, part of the coastal plain.

Uchupi (1970) in a study covering the full east coast of the U.S. provided some basic information of the structural setting and gross stratigraphy of the upper segment of the shelf but the study's methods did not allow any detailed interpretation of the Late Tertiary and Quaternary strata. Richards (1967, 1974) described the Fort Monroe High, a bedrock structural high that crests near the mouth of Chesapeake Bay, and adjacent lows, or troughs, the deepest portions of which are beneath Ocean City, Maryland and Cape Hatteras, North Carolina. Richards (1967, 1974) reported that a well at Fort Monroe, Virginia reached crystalline rocks at -681 m (-2,234 ft) whereas crystalline rocks were encountered at -3,041 m (-9,978 ft) beneath Cape Hatteras and at -2,350 m (-7,710 ft) at Ocean City. More recent work (Poag, 1997, among others) indicates that the well at Fort Monroe is within the Eocene Chesapeake Bay Impact Crater and may yield an anomalous depth for basement.

Bayer and Milici (1987) studied the deeper sedimentary structure underlying the continental shelf off Virginia with an emphasis on the potential for petroleum exploration. They reported a set of diapirs in the strata underlying the 200 m isobath, roughly along what they considered the "paleoshelf edge." They also noted a discontinuous, Mesozoic reef beneath the shelf edge, roughly in line with the diapirs and two synrift basins about 40 km offshore. They concluded that "the probability of the Virginia continental margin containing commercially recoverable quantities of oil and/or gas [ranges] from fair (moderate) to poor (low)."

Because the Chesapeake Bay Impact Crater was discovered only in the early 1990s (Powars *et al.*, 1993), it is only recently that realization of the influence the crater has on the more recent geology has grown. In some cases the influence is only inferential, whereas in others the existence of the crater provides reasonable explanations for features noted by earlier researchers. Perhaps differential compaction between the crater fill and the surrounding, undisturbed strata helped define the arcuate paleoshoreline that became the Suffolk Scarp. Johnson *et al.* (1998) remarked that the few earthquakes that have occurred in the Virginia coastal plain were located very close to the outer rim of the crater.

COASTAL GEOMORPHOLOGY AND PROCESSES

Swift *et al.* (2003) and Parsons *et al.* (2003) synthesized previous studies to formulate the following "rules for bedload sediment dispersal on the Virginia Coast."

- At time scales of fifty years of greater, the entire shoreface in transgressive sectors of the Middle Atlantic Bight is undergoing landward retreat.
- Net along-shore sediment flux in the surf zone (0 to 3 m water depth) of the Middle Atlantic Bight is southward to southwestward and parallel to isobaths, with the exception of local reversals south of inlets and major estuaries.
- Net across-shelf transport averaged over decadal time intervals, to include major storms, has a component directed onshore, reflecting episodic storm washover of the barrier islands.
- Grain size on the beaches of coastal compartments of the Middle Atlantic Bight fines downdrift to the south.
- Sand on the inner Delmarva shelf (landward of the 20-m isobath) fines alongshore to the southwest, down the transport path, from 1.8 phi (medium sand) at Chincoteague Shoals to 2.4 phi (fine sand) at Smith Island.
- Sediment transport on the middle and lower shoreface (water depths of 5 to 20 m) is directed southward and offshore, and is oblique to isobaths.
- The intensity of storm-driven, across-shelf sediment flux diminishes seaward as a function of increasing water depth and decreasing intensity of wave orbital motion during peak storm flows.
- Sediment transport over storm-generated shoreface sand ridges in the Middle Atlantic Bight is typically south and offshore, moving obliquely across the crests.
- The troughs between the shoreface ridges undergo erosion, as do the up-current flanks of the ridges, while the crests and down-current flanks aggrade, as the ridges migrate slowly downdrift.

In the present coastal zone of the Mid-Atlantic, nearly all coarse clastic material (sand and gravel) carried by rivers draining the Piedmont and Appalachians is deposited at the head of the tidal estuarine portion of the river (Meade, 1969, 1972). Finer-grained particles (silt and clay) discharged by the rivers generally are held in suspension until flocculation and rapid deposition occur in the zone of maximum turbidity, which coincides with the initial increase of dissolved ions (salinity) in the upper reach of the estuary (Nichols, 1972; Gibbs, 1987; Nichols *et al.*, 1991). Consequently, nearly all of the sand moved about in the coastal zone and on the shelf has been eroded and reworked from previously deposited sediments. Swift *et al.* (1971) applied the term *palimpsest* in describing the sediment-starved Mid-Atlantic shelf.

Sediments

Shepard (1932) characterized the surficial sediments of the inner shelf near the Virginia-North Carolina border as "shells, sand & gravel" and "gravel," and "shells & sand" near the mouth of Chesapeake Bay. Based on a suite of grab samples collected on an 18 km (10 n mi) grid, Milliman (1972) and Milliman *et al.* (1972) reported a patch of "subarkosic to arkosic, fine-grained sediments" on the inner shelf roughly between the Virginia-North Carolina border and Chincoteague Island. In contrast, most of the Virginia shelf to the 60 m (200 ft) isobath is covered with "arkosic to subarkosic sands." They characterized the finer-grained sediments closer to shore as modern "nearshore (fluvial?) deposits" and the sands as relict "fluvial."

Milliman *et al.* (1972) also described aspects of the heavy mineral distribution in shelf sands. They showed that garnet, which is derived from erosion of crystalline rocks and is resistant to weathering, generally constitutes sixteen to thirty percent of the heavy mineral fraction of the shelf sand; two large patches, one generally offshore of the Chesapeake Bay mouth and extending to about the 60 m (200 ft) contour and the other being a large section running from the Delaware-Maryland shore to roughly the 60 m (200 ft) contour offshore of the Maryland-Virginia border, have garnet concentrations as high as forty-five percent. The garnet fraction was lower at the

Chesapeake Bay mouth and on the outer shelf near the head of the Norfolk Canyon. Amato (1994) reviewed and summarized the earlier works and provided maps of the distribution of sediment types.

Hobbs (1997) reported on the grain-size characteristics of a set of almost 400 grab samples collected between the Chesapeake Bay mouth and the Virginia-North Carolina border from the shoreline east about 60 km to approximately 75°14' W longitude, with maximum water depths of 38 m (126 ft) (Figure 2). The dense sample grid provided more detail on the distribution of surface sediment types than the earlier, reconnaissance studies. Sands and granules are the dominant surface sediment. There are two nearshore pockets with a slightly elevated fine-grained component. The first, immediately south of the Chesapeake Bay mouth likely results from suspended-sediment transport out of the Bay. The second is farther south, in the vicinity of False Cape, and might represent an outcrop on the shelf of relict back-barrier lagoonal sediments. Other small areas of fine-grained sediments on the southern portion of Virginia's continental shelf are of uncertain origin but might be controlled by subtle topographic variations or result from outcrops of older strata.

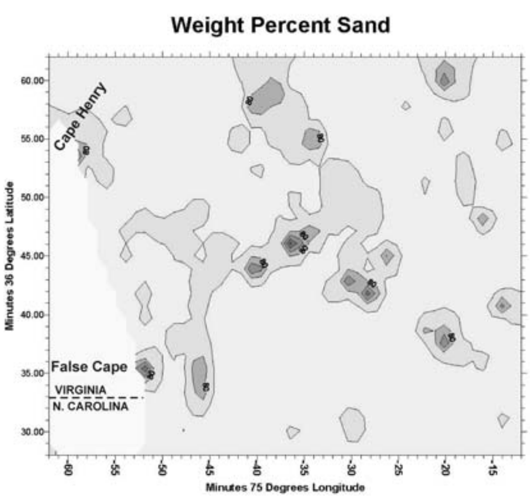


Figure 2: Weight percent sand from 380 samples offshore of southeastern Virginia (Hobbs, 1997).

Swift *et al.* (1971) considered the heavy mineral suites in the 0.085 to 0.177 mm (fine to very fine sand) size fraction along the coast between Cape Henry, Virginia and Cape Hatteras, North Carolina. They found "three well-defined heavy mineral provinces" across the shore and a less distinct alongshore variation. Two zones dominated by the amphibole-garnet-kyanite suite on the beach and offshore are separated by an amphibole-epidote-kyanite suite in the nearshore. The three suites "correspond to three well-defined grain-size provinces generated by the contrasting hydraulic regimes of the surf, zone of shoaling waves, and the shelf floor." The major longshore trends are southward, with an increase in garnet and opaque minerals and a corresponding decrease in amphibole.

While the studies by Milliman (1972) and Milliman *et al.* (1972) on the petrology of the heavy minerals in the sand fraction were primarily sedimentological, Grosz and Eskowitz (1983) considered the economic potential of heavy-mineral deposits on the continental shelf. Other studies (Berquist and Hobbs, 1986, 1988a, b, 1989; Ozalpasan, 1989; Berquist, 1990; Berquist *et al.*, 1990; Grosz *et al.*, 1990; Dydak, 1991) specifically addressed the economic heavy minerals of the inner shelf, generally within 10 km (6 mi) of the Virginia shoreline. According to Berquist *et al.* (1990), at least four areas satisfy the criteria of Garner (1978) for threshold levels of economic heavy minerals: off Hog Island, and off Smith Island, in the southern half of the Virginia Eastern Shore; off Virginia Beach just south of the Chesapeake Bay mouth; and off False Cape near the Virginia-North Carolina border. The minerals of potential interest on the Virginia shelf are the titanium minerals ilmenite, leucoxene, and rutile, along with zircon and monazite.

The Shore

The coast of Virginia consists of the ocean shore of part of the Delmarva Peninsula and the ocean shore of southeastern Virginia, which is entirely within the City of Virginia Beach. As such, Virginia's coast is part of two coastal compartments: 1) the full reach of the Delmarva Peninsula from Cape Henlopen, Delaware to Cape Charles, Virginia and 2) the reach extending south from Cape Henry across the Virginia–North Carolina border to Cape Lookout. The approximate distance from Cape Henry to the state line is 40 km (26 mi) and from Cape Charles to the Maryland boundary is 122 km (76 mi). Essentially the coast line is the outer limit of the coastal plain, which stretches nearly 125 km (75 mi) from the Fall Zone along the eastern edge of the Piedmont to the Atlantic Ocean.

Prior to the onset of the glacial episodes, the major rivers in the region, the Delaware, Susquehanna, Potomac, Rappahannock, York, and James Rivers, flowed generally southeast from the Piedmont across the upper coastal plain to the Atlantic Ocean. The growth of the Delmarva Peninsula (Mixon, 1985), first as a reworking of deltaic deposits, then as a progression of major spits, each major spit forming in response to an interglacial sea-level high (Oertel and Foyle, 1995; Hobbs, 2004; Oertel and Overman, 2004) deflected the courses of the rivers and created Chesapeake Bay (Figure 3). The paths of several filled paleochannels (Colman *et al.*, 1990, among others) as they pass under the Delmarva Peninsula and cross the continental shelf, along with the high-sea level deposits on the intervening areas, constitute the antecedent, or framework, geology of the present coastal zone and continental shelf.

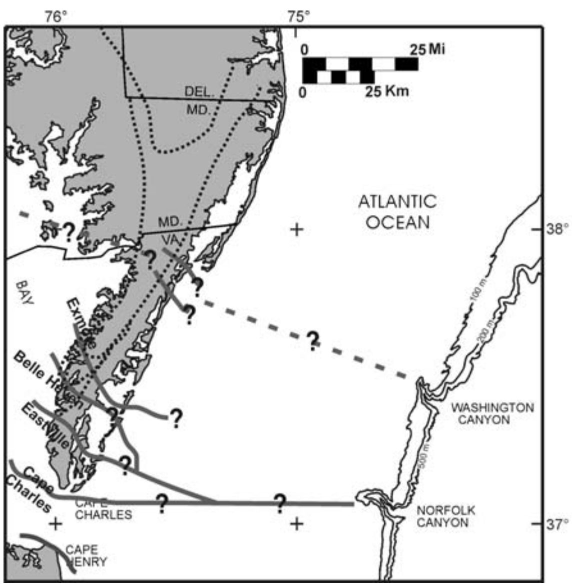


Figure 3: Southward extension through the Pleistocene of the southern Delmarva Peninsula as a large-scale prograding spit, and interaction with valleys of the primary rivers flowing across the Virginia and Maryland coastal plain. (Adapted from Hobbs, 2004.)

The coastal plain has a series of stairsteps that are broad coast-parallel plains, or terraces, each bounded on the landward side by a scarp with a sharp increase of elevation. The scarps represent the previous position of erosional shorelines, and the terraces would have been the floor of the shelf stretching seaward. In many cases, the terraces were reoccupied by later highstands of sea level that left barrier islands stranded as paleo-shorelines. The entire system of scarps and terraces was created by a series of successively lower highstands of sea level that cut into and planed off existing sediments. The older scarp and terrace sets were preserved because of the slow but continuous uplift of the Piedmont and coastal plain.

GEOGRAPHY OF THE VIRGINIA AND DELMARVA COAST

Major Rivers in the Region

One of the central themes in the development of the Virginia coast is the alternation between processes operating during sea-level highstands and those during lowstands. The major rivers of the region have been integral in the geologic history by episodically delivering large volumes of coarse sediment that built up the coastal plain and shelf, and by excavating valleys that generally trend southeast across the continental margin toward the offshore submarine canyons. The channels and broader valleys incised by the rivers and their tributary drainage networks during lowstands provided accommodation space that would be flooded by rising sea level to form estuaries during highstands. The rivers crossing the Virginia coastal plain also interacted with the basin created by the Chesapeake Bay Impact Crater, and in large part controlled the evolution of the Virginia Eastern Shore prograding south as a large-scale spit through the Pleistocene.

As noted above, the major rivers that drain the Appalachian highlands, the Delaware, Susquehanna, Potomac, and James Rivers, have remained in essentially the same positions for at least the past 10 My (Pazzaglia, 1993). Fluvial sands and gravels from these rivers cap the eroded hills of the Piedmont and rim the western edge of the coastal plain along the Fall Zone (Mixon *et al.*, 1989). Although continental glaciers did not extend as far south as Maryland and Virginia, the Delaware and Susquehanna Rivers received outwash from the distal ends of melting glaciers in central Pennsylvania and northern New Jersey. As these sediment-choked rivers reached the break in slope at the inner coastal plain, they spread out as braided-river systems that deposited great sheets of coarse sand and gravel across central Delmarva. Similar sheets of coarse clastics from the ancestral Potomac River built out the peninsula that is now Southern Maryland, between the modern Potomac and Patuxent Rivers (McCartan *et al.*, 1990).

The intermediate rivers on the Virginia mainland, the Rappahannock and York Rivers, drain the Piedmont and make short excursions into the front of the Blue Ridge. Geologic evidence suggests that these smaller rivers also have been in approximately the same positions at least since the late Pliocene (about the last 2 My) (Ramsey, 1992; among others). Both of these rivers presently are flooded as tidal estuaries, with the head of tides close to the Fall Zone. However, their most significant contribution to the evolution of the Virginia coast was carving the valleys that would later provide outlet courses for the combined Susquehanna-Potomac drainage as it was displaced southward through the Pleistocene (Figure 3) (Oertel and Foyle, 1995; Hobbs, 2004).

Atlantic Coast of the Delmarva Peninsula

Fisher's (1967) model of coastal compartments as modified by Oertel and Kraft (1994), characterizes the geomorphology of the Atlantic coast of the Delmarva Peninsula in terms of supply of sand to the shoreline by erosion and redistribution by transport processes. This review of the coastal geomorphology follows the coastal compartment model and relates the geomorphic changes along the coast to setting and processes.

A brief history of the Pliocene to Pleistocene evolution of the Delmarva Peninsula is relevant to the modern coastal zone. The core of the peninsula developed as a series of river deltas and braided-river outwash plains that deposited coarse sediments (sand and gravel, with some cobbles and boulders) (Ramsey, 1992; Pazzaglia, 1993). Age control for the major depositional events is generally poor because these deposits have few fossils for biostratigraphy or materials for radiometric or other geochemical dating.

The principal surficial or near-surface stratigraphic units on Delmarva are the lower Pliocene (or upper Miocene) Pensauken Formation, which covers much of the land surface of the Maryland Eastern Shore south to the Choptank River (Owens and Denny, 1979); the upper Pliocene Beaverdam Sand, which occurs in south central Delmarva, primarily in Wicomico County, Maryland and Sussex County, Delaware (Owens and Denny, 1979; Ramsey and Schenck, 1990; Andres and Ramsey, 1999); and the lower Pleistocene Columbia Formation, which mantles about two-thirds of Delaware (Groot *et al.*, 1995; Groot and Jordan, 1999; Ramsey, 1999).

Since the last major depositional event in the early Pleistocene, the high-energy Atlantic shoreline has cut into these deltaic deposits during highstands, and the Delaware and Susquehanna Rivers have incised during lowstands to carve out the basins for the Delaware and Chesapeake Bays, respectively. Sand eroded by vigorous wave attack was transported south and southwest by the regional longshore transport system that operated basically the same as the modern system. Deposition downdrift produced the series of large-scale spits that prograded into the basin created by the Chesapeake Bay Impact Crater and formed the Virginia Eastern Shore.

Assateague Island and North

From Cape Henlopen, Delaware, to the southern tip of Assateague Island in Virginia, the Delmarva coast essentially is one continuous, long, narrow barrier island interrupted by only two inlets, both of which are maintained: Indian River, Delaware, and Ocean City, Maryland. The northern 30 kilometers (18.5 miles) of this coastline comprise the eroding headland and cuspate foreland elements of the Delmarva coastal compartment (Oertel and Kraft, 1994). This shoreline derives sand from erosion of the shoreface and Pleistocene headlands at Rehoboth Beach and Bethany Beach, Delaware. Shoreface ravinement is the primary erosional process, as scour by storm waves and longshore currents bevels Holocene back-barrier deposits exposed in the shoreface and excavates shallow Pleistocene sediments to 11 to 13 m below present sea level (Kraft, 1971).

Sand in the littoral transport system in these northern sections moves north, counter to the regional transport direction. Longshore transport diverges in a nodal zone along the Delaware-Maryland coast. The actual boundary between northward and southward flow varies between years as net transport balances the effects of winter northeaster storms and summer southeasterly breezes. Consequently, the point of transport divergence migrates between South Bethany and the easternmost point on the Delmarva coast near the Delaware-Maryland state line at Fenwick Island (Dalrymple and Mann, 1985). The northward flow of sand along the Delaware coast is driven both by the north-northwest orientation of the shoreline (rotated west from the northeast orientation of the Maryland coast) and the wave-energy gradient into Delaware Bay. The flow of sand in the littoral system has created the baymouth barrier or barrier spit that closed two incised valleys to form Rehoboth and Indian River Bays, and to build the cuspate foreland of Cape Henlopen, where the shore abruptly turns westward into Delaware Bay (Kraft *et al.*, 1978).

From the Delaware-Maryland state line, regional longshore transport is to the south all the way to Cape Charles at the mouth of Chesapeake Bay. Local reversals of transport occur immediately downdrift of tidal inlets. The typical scale of downdrift effects by an inlet is about 6 km (4 mi) (Fenster and Dolan, 1996) and varies with the size and stability of the inlet and associated ebb-tidal delta.

Fenwick and Assateague Islands extend south 73 km (45 mi) from the Delaware-Maryland border to the southern tip of Assateague Island in Virginia. The resort towns of Fenwick, Delaware and Ocean City, Maryland occupy Fenwick Island, which is heavily developed and hosts upwards of 300,000 visitors on a typical summer weekend. In contrast, Assateague Island is minimally developed, with extensive pristine areas and restricted access to much of the island. The Maryland section of Assateague Island has been owned and managed by the National Park Service since 1965 as Assateague Island National Seashore, with a Maryland State Park as an inholding. The Virginia section is managed principally by the U.S. Fish and Wildlife Service as Chincoteague National Wildlife Refuge.

Assateague Island is bounded to the north by Ocean City Inlet, which was cut by a major hurricane in 1933 and stabilized with jetties soon after by the U.S. Army Corps of Engineers. The jetties and, more importantly, the flood- and ebb-tidal deltas of the inlet have trapped a significant volume of sand that otherwise would have been transported to Assateague Island (Kraus, 2000). As a consequence, the northern 5-km (3-mi) section of Assateague has migrated landward a full island width, more than 500 m (1,600 ft), and experiences nearly complete overwash during major storms. The arc of erosion created by the inlet extends nearly 16.2 km (9.7 mi) south on Assateague Island (Galgano, 1998).

Central Assateague Island, south to Green Run Bay near the Maryland-Virginia border, is a typical wave-dominated barrier island, generally between 500 m and 1 km (0.3 to 0.6 mi) wide with broad overwash flats extending into the back-barrier lagoon of Chincoteague and Sinepuxent Bays. Higher-elevation sections of the island are dissected by the low-lying traces of previous tidal inlets, some of which were open for a few decades during the past 150 years (McBride, 1999). One of the more prominent inlets is Green Run Inlet, which was open through much of the early to mid 1800s, but shallowed, filled, and finally closed between 1880 and 1900.

From Green Run Bay south, the modern Assateague Island is offset about 1 km (0.6 mi) seaward of Pope Island and Chincoteague Island, which have been interpreted as late Holocene precursors to the modern barrier island (Halsey, 1978, 1979). Assateague Bay Inlet cut through Assateague Island seaward of the north end of Chincoteague Island during the Colonial period (1600s and 1700s) (Amrhein, 1986). By about 1800, this inlet was closed by the first of a set of spits extending south and wrapping around the eastern side of Chincoteague Island (Goettle, 1978). The original Assateague Lighthouse was constructed in 1833 on the highest-elevation and most prominent of these spits; at the time, this was the ocean shoreline. Today the second lighthouse, built on the same site in 1867, is 3 km (1.8 mi) inland from the beach.

The southern end of Assateague Island is the terminus of the regional longshore sediment transport system moving sand south nearly 75 km (46 mi) from the Bethany headland (Oertel and Kraft, 1994). The informally named Toms Cove Hook is the spit complex of south Assateague, with Fishing Point as the distal tip. This spit extended south nearly 6 km from a cuspate foreland that was the end of the island in the mid 1800s (Field and Duane, 1976). Initial extension of the spit was south-southwest onto the existing bathymetric high of Ship Shoal. Historical charts and maps indicate that by approximately 1890, the leading edge of the spit encroached on a deep trough between Ship Shoal and Turners Lump and turned sharply to the west. Continued progradation to the west-northwest during the 1900s partially enclosed a relatively deeper basin west of Ship Shoal to form Toms Cove.

The net rate of sand transport to this spit complex has been debated, with estimated annual accumulation rates within the combined Toms Cove Hook and Chincoteague Inlet system ranging from 165,000 m^3 per year to 1,100,000 m^3 per year (Finkelstein, 1983; Headland *et al.* 1987). The closure of the historic Green Run and Pope Island inlets (and possible other unnamed inlets) and subsequent reworking of their respective ebb-tidal deltas in the late 1800s *may* have contributed to the spit's rapid growth (Halsey, 1978; Demarest and Leatherman, 1985; McBride, 1999).

Chincoteague Inlet has one of the longest histories of dredging to maintain a navigable channel of any seaport entrance along the U.S. East Coast. In recounting the recent activity, Morang *et al.* (2006) stated that the federal navigation project at Chincoteague Inlet is dredged to 3.7 m depth with overdredging on the ebb shoal of about 1 m. The quantity of material dredged during eight projects between 1995 and 2006 varied between about 53,500 and 94,000 m^3 (70,000 and 122,900 yd^3), excepting a small effort of about 9,500 m^3 (12,500 yd^3) in 2005. The total quantity of material removed during that period was almost 473,000 m^3 (620,000 yd^3).

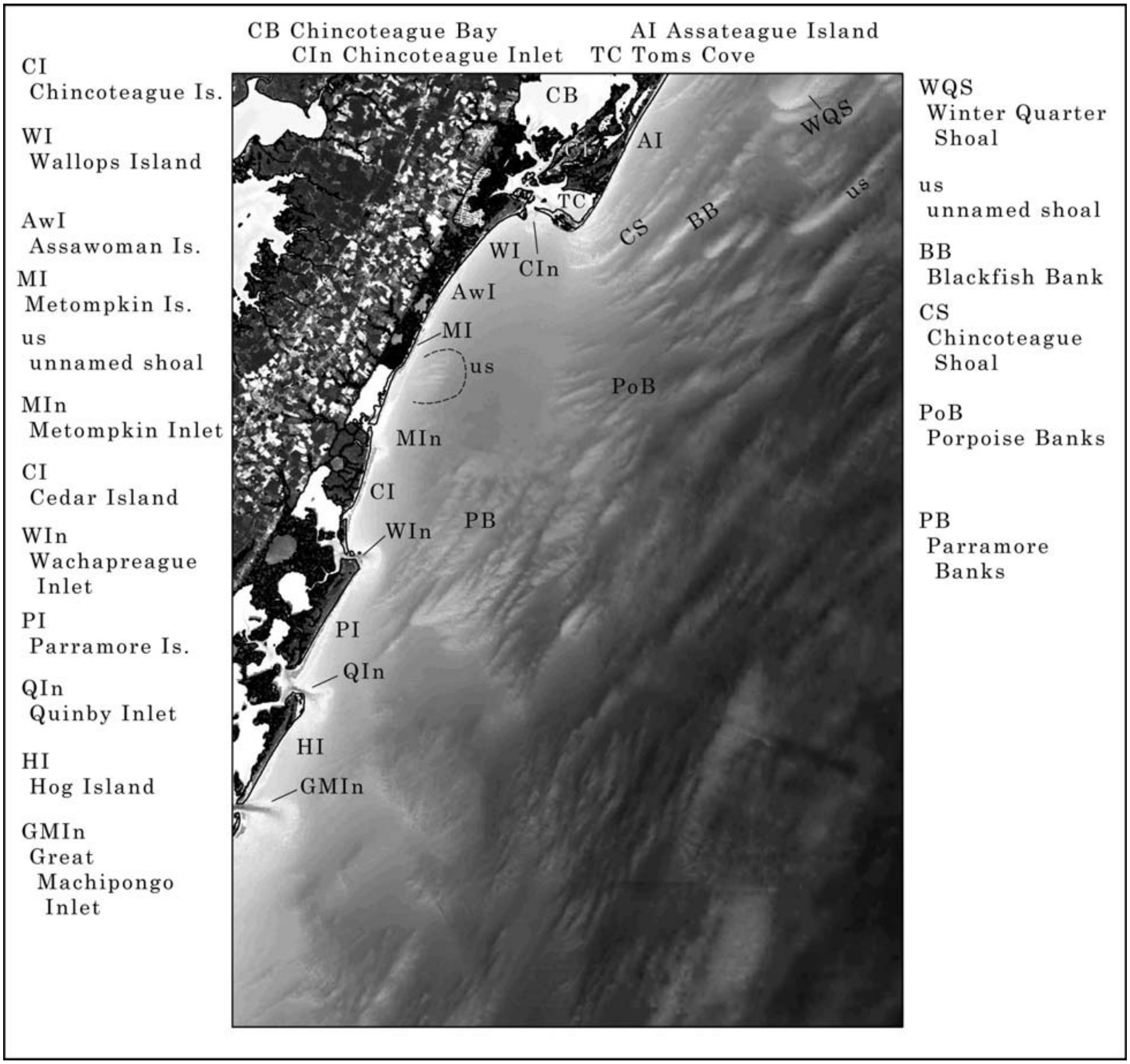


Figure 4: Coastal zone and shaded-relief bathymetry of the inner shelf for the Delmarva coast from southern Assateague Island to Hog Island, showing the landward offset of barrier islands forming the Chincoteague Bight.

The nearshore shelf surface off Assateague Island is floored by a discontinuous sheet of medium to fine sand, molded into shore-oblique ridges and swales that are spaced 2-4 km (1-2.5 mi) apart and extend kilometers to tens of kilometers (Swift *et al.*, 2003). These sand ridges generally are oriented southwest-northeast with a maximum relief of 5-10 m (15-35 ft). Shoreface ridges are the most significant source of coarse sediment on the inner shelf (Toscano and York, 1992), and the largest number and highest density of ridges along the East Coast occur off Assateague Island (McBride and Moslow, 1991).

A prominent erosional surface generally underlies the actively transported sands of the shoreface ridges (Toscano and York 1992). Seismic data show an inclined, planar to hyperbolic surface with a fairly uniform geometry, except for outcropping platforms where sediment is being exhumed from fine-grained back-barrier facies. The sand ridges are post-transgressive expressions of modern shoreface processes, in particular a response to wave action and alongshelf currents generated by storms (Swift and Field, 1981; Toscano and York, 1992; Snedden and Dalrymple 1998). These energetic events mobilize sand in the shoreface, where waves may transport it across the barrier island to create washover fans, or downwelling currents may sweep it off the upper and middle shoreface into deeper water offshore (Parsons *et al.*, 2003).

Wells (1994) and Conkwright and Gast (1995) found that many shore-attached shoals were relatively narrow bodies of fine sand overlying a core of finer sediments. McBride and Moslow (1991) hypothesized that the ridges developed from ebb-tidal deltas and owe their oblique orientation to inlet migration and shoreface retreat. The sand ridges disappear abruptly in a comparatively featureless basin south of Fishing Point.

The Chincoteague Bight

One of the major geomorphic transitions along the Virginia coast occurs immediately south of Chincoteague Inlet, where Wallops Island is offset nearly 6 km (3.6 mi) west of Assateague Island (Figure 4). The next three islands south – Assawoman, Metompkin, and Cedar Islands – are offset as much as 10 km (6 mi) west of a line drawn between Assateague and Parramore Islands, to form an arcuate embayment of the coast that has been named the Chincoteague Bight (Oertel *et al.*, 2007). The relatively abrupt transition at Chincoteague Inlet marks the boundary between the wave-dominated islands to the north and the mixed-energy, "drumstick" islands with large ebb-tidal deltas to the south (Finkelstein, 1983; McBride and Moslow, 1991; Oertel and Kraft, 1994) (Figures 1 and 4). The morphologic change also coincides with a transition to finer sand to the south (Swift, 1975).

Wallops Island, the northernmost barrier in the Chincoteague Bight, has the classic bulbous updrift end and narrow downdrift spit of a mixed-energy island (Hayes, 1979, 1980). Wave refraction around Chincoteague Shoals, Fishing Point, and the ebb-tidal delta of Chincoteague Inlet drives the bypassing of sand around the outer edge of the delta. The re-attachment of swash bars migrating around the delta, and a local reversal in longshore transport to the north result in accretion to the northern end of Wallops Island. However, the process with the greater effect on the coastline is the capture of a huge volume of sand by the flood-tidal delta landward of the inlet. The 12-km by 6-km (7.4-mi by 3.7-mi) section of the back-barrier lagoon behind Chincoteague and Wallops Islands is filled by this extensive tidal delta and marsh complex.

The interruption of regional longshore transport by Chincoteague Shoals and Chincoteague Inlet is a primary factor contributing to the starvation and rapid retreat (2-9 m/yr, or 6.5-30 ft/yr) (Rice and Leatherman, 1983; Kraus and Galgano, 2001) of the 35 km (20 mi) arc from Wallops Island to Cedar Island. The southern section of Wallops Island, which is the most stable island of the group, retreated about 400 m from 1857 to 1994 for an average rate of 2.9 m/yr (9.5 ft/yr) (Morang *et al.*, 2006). Locally near Assawoman Inlet, between Wallops and Assawoman Islands, the rate increases to 3.7 to 5.5 m/yr (12 to 18 ft/yr), and Assawoman Island itself retreated about 5 m/year (16.5 ft/yr) between 1911 and 1994 (Morang *et al.*, 2006). The present surfaces of Assawoman and Metompkin Islands are extremely low, and are essentially an amalgamation of thin overwash fans migrating across the back-barrier marsh. It is often difficult to define the shoreline of either island on aerial photographs. In other words, because of sediment starvation and rapid transgression, at times these barrier islands essentially cease to exist.

Figure 5: Mosaic of the Virginia Eastern Shore from photos taken from the International Space Station. (NASA images ISS004-E-11554 and 11555.)

The inner shelf of the Chincoteague Bight also is anomalous along the Delmarva coast. The shoreface ridges and offshore shoals prominent off Assateague Island end abruptly southeast of Toms Cove Hook (Figure 4). In contrast, the inner shelf directly off Metompkin and Assawoman Islands is smooth, featureless, and relatively deep. Along the southern margin of the Bight, the sand shoals of Parramore Banks create some bathymetric relief. Otherwise, within the Bight the only shoals are the thin stringers of Porpoise Banks nearly 15 km (9 mi) offshore in water 16-18 m (52-59 ft) deep, and an unusual, unnamed, set of nearshore shoals 2 km (1.2 mi) off Metompkin Island (Figure 4).

In a study of the inner shelf adjacent to Cedar Island, immediately north of Wachapreague Inlet, Wright and Trembanis (2003) and Trembanis (2004) noted that this section of the shoreface is substantially deficient in potentially mobile sand. Their mapping of the seafloor with high- resolution side-scan sonar revealed considerable variability, and showed areas of exposed, relict, Holocene marsh peat and filled back-barrier channels. They attributed the variability to localized deposits of shell and a thin, mobile, sheet of sand overlying the relict back-barrier deposits. The sharp transitions between bottom types coincided with increased roughness of as much as an order of magnitude between sand and the exposed shell or peat, which can substantially affect the bottom boundary-layer thickness, bottom drag on currents, and the vertical concentration of suspended sediment (Masden *et al.*, 1993; Wright,1993).

The Virginia Barrier Islands

From Parramore Island south to Cape Charles, the southern tip of the Delmarva Peninsula, the barriers are the short, stubby islands typical of a mixed-energy coast (Figure 5). The Virginia barriers sit 6-8 km (3.7-5 miles) to as much as 12 km (7.4 miles) away from the mainland. The back-barrier is a complex of tributary tidal channels, shallow subtidal to intertidal mudflats, extensive tidal marshes, and open bays. The larger inlets, such as Great Machipongo and Sand Shoals Inlets, have deeply scoured channels 20 to nearly 25 m (65 to 82 ft) deep, 5 km (3 mi) long, and 500 m (more than 0.25 mi) wide. The ebb-tidal deltas extend seaward 5-6 km (3 to 3.7 mi), and as much as 8-9 km (5-5.5 mi) alongshore, with a pronounced downdrift (southward) skew.

Sand transport dynamics in this coastal compartment are dominated by the tidal inlets and coast-perpendicular tidal flow. The inlets and associated ebb-tidal deltas trap and hold sand from the upper shoreface. Even so, sand in the middle to lower shoreface (roughly 10 to 20 m, or 33 to 65 ft, water depth) continues to move south toward the mouth of Chesapeake Bay (Swift *et al.*, 2003). Overall net sand transport in the shoreface is slowed dramatically relative to that in the wave-dominated compartment to the north.

Wave refraction around the outer edge of the ebb-tidal deltas, combined with bypassing via the migration of bars or shoals attached to the seaward apron of the deltas, produces a zone downdrift of the inlet in which longshore transport reverses (flowing to the north) and additional sand is added to the updrift end of the island. By this process, most of the Virginia barrier islands have a bulbous updrift end that is relatively wide with multiple beach ridges; in contrast, the downdrift end typically is a narrow, low, prograding spit that is unstable and frequently overwashed during storms. Most of the Virginia barriers also exhibit downdrift offsets (Hayes *et al.*, 1970; Hayes, 1979, 1980); the southern island of two framing an inlet will be as much as 1 km (0.6 mi) seaward of the northern island. Because of the differences in width and stability of the island sections, they tend to rotate in response to transgression rather than undergoing parallel retreat (Rice and Leatherman, 1983). The maximum distances of inlet influence extend to 6.8 km (4 mi) updrift and 5.4 km (3.3 mi) downdrift of inlets along the Virginia barrier islands (Fenster and Dolan, 1996).

Rice *et al.* (1976) and Rice and Leatherman (1983) reviewed and compiled information on the Virginia barrier islands for The Nature Conservancy. They included information about shoreline changes, physical processes, geomorphology, and geologic history of the individual islands as a guide for making decisions about future management and utilization of the island chain. The report provided complete discussions of each island and addressed the interrelationships among the islands, inlets, and lagoons.

Leatherman *et al.* (1982) and Rice and Leatherman (1983) categorized the Virginia Barrier Islands into three geomorphic groups based on the response to transgression: 1) A northern group that extends from Assateague Island to Cedar Island (within the Chincoteague Bight), characterized by parallel beach retreat; 2) a middle group including Parramore, Hog, Rogue, Cobb, and Wreck Islands, characterized by "rotational instability;" and 3) a southern group of Ship Shoal, Myrtle Island, and Smith Island, characterized by non-parallel beach retreat. They attributed the variations in geomorphology to differing sediment supply, underlying Pleistocene topography, and tectonic movements. It is worth noting that this paper was written prior to the discovery of the Eocene

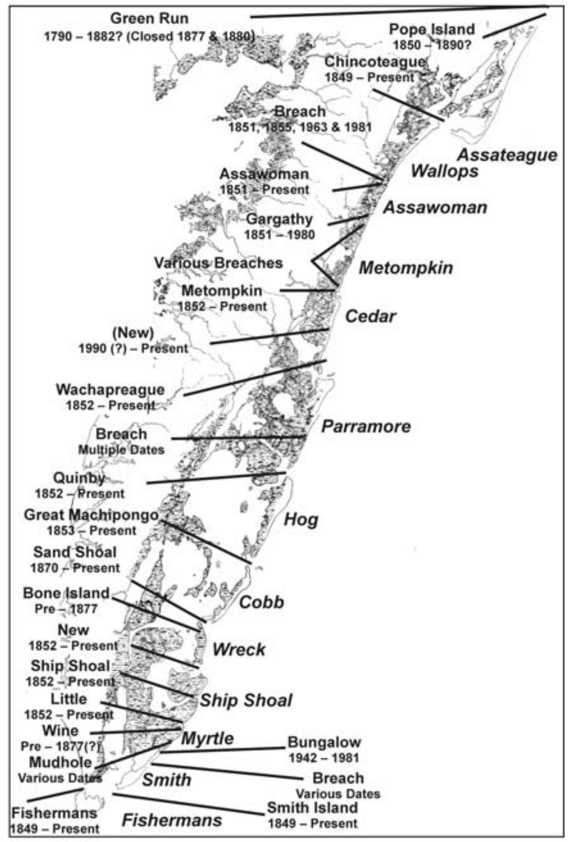


Figure 6: Location of present and historic inlets in the Virginia barrier island chain. Initial dates refer to dates of the first chart on which inlet was named. Names of barrier islands are in italics. Modified from McBride (1999).

Chesapeake Bay Impact Crater. The Inner (or "Peak") Ring of the crater neatly marks the separation between the middle and southern groups. The Outer Ring of the crater is only slightly south of Wachapreague Inlet, which separates the middle and northern groups (Poag, 1997, among others). It is likely that differential compaction and other local tectonics associated with the crater account for the "tectonics" cited by Leatherman *et al.* (1982). The filled Eastville and Exmore paleochannels cut by the Susquehanna-Potomac River system (Colman *et al.*, 1990) also underlie the transitions between island groups.

McBride (1999) cataloged the present and historic tidal inlets along the entire Delmarva Peninsula (Figure 6). He distinguished two primary modes of barrier-island response. In the first case, inlets along wave-dominated barriers tend to open by island breaching during storms. They then rapidly migrate laterally in a southerly direction in response to net longshore transport, and close after a few years to a few decades (i.e., these inlets are short-lived). For the second mode, inlets along tide-dominated barriers tend to be nonmigratory (i.e., they oscillate in place) and remain open much longer. The main factors attributed by McBride (1999) that control inlet behavior and distribution are tidal range, wave energy (height) and direction (driving net littoral drift), tidal prism, storm magnitude and recurrence, and antecedent topography. Of these factors, wave and tide energy were determined to be most important.

Figure 7: Erosion on Parramore Island, Virginia. The rapidly retreating shoreline is cutting into the island's maritime forest. Photo was taken in August 2006.

McBride and Vidal (2001) characterized the ocean shore of Parramore Island as having three distinct zones. The northernmost zone is influenced by Wachapreague Inlet and the ebb-tidal delta, and the shoreline has retreated since 1980. The second zone is the north-central segment of the island, which is characterized by truncation of tree-lined beach ridges (Figure 7). Although historically fairly stable, the central zone has experienced accelerating retreat since 1962 (4.6 m/yr from 1959/62-1980, 8.3 m/yr from 1980-1998, and 9.0 m/yr from 1998-2000). Wikel (2005) used LIDAR data from 1997 and 2004 to determine the average annual rates of shoreline retreat of 12 m/yr and 10 m/yr for the northern and middle sections of the island, respectively. The third zone is the southern half of the island, which is a low, narrow spit dominated by washover processes. The rapid retreat of the southern section – 7.8 m/yr from 1871 to 1959/62, 6.4 m/yr from 1959/62 to 1980, 12.8 m/yr from 1980 to 1998, and 15.6 m/yr from 1998 to 2000 – contributes to the clockwise rotation of the island. McBride and Vidal (2001) attributed the accelerating erosion rate to four factors: 1) extension of the arc of erosion south of Fishing Point (the southern end of Assateague Island); 2) impact of the 1962 Ash Wednesday storm; 3) morphodynamics of the large ebb-tidal delta at Wachapreague Inlet; and 4) continuing relative sea-level rise.

The Virginia Coast Reserve / Long Term Ecological Research Site (VCR/LTER) on the barrier and back-barrier system of the southern Delmarva segment has been the focus of numerous integrated geological and ecological studies. The VCR/LTER research activities "focus on the mosaic of transitions and steady-state systems that comprise the barrier-island/lagoon/mainland landscape of the Eastern Shore of Virginia" (Virginia Coast Reserve, 2007). Primary study sites are located on Hog Island, Parramore Island, and mainland marshes near Nassawadox, Virginia.

In a study of the physical factors influencing growth or loss of tidal marshes landward of the Virginia barrier islands, Oertel *et al.* (1989) concluded that the elevation of the marsh surface with respect to the elevation of mid-tide, the rate of relative sea-level rise, and sediment outwash from the mainland work together to control the net rate of marsh accretion. They emphasized the need to differentiate between marsh accretion resulting from short-term events and from long-term accumulation of suspended sediment.

Christiansen *et al.* (2000) characterized the processes that control sedimentation on the surface of back-barrier marshes. The combination of flocculation of suspended particles and dramatic reduction of turbulence facilitate deposition. The aggregation of fine particles into flocs accounted for about 75 percent of the inorganic material deposited within 8 m of the tidal creek. The larger particles that result from flocculation have appreciably higher settling velocities that allow deposition to the marsh surface. Low flow velocities (<1 cm/s) and correspondingly low shear stress across the marsh surface initiate particle settling and limit resuspension. They observed that concentrations of suspended sediment at the marsh edge are greater on rising tides than falling, which supports the interpretation of sediment accumulation on the marsh surface. Sediment concentrations decrease with distance from the marsh creek because deposition out of the flow reduces suspended sediment concentration; this process also partially explains the development of levees along the banks of marsh creeks.

Erwin *et al.* (2004) examined multi-decade changes in the morphology of salt marshes at five locations, three of which are in the back-barrier lagoons of the Virginia barrier islands. Not surprisingly, marsh area was lost at sites adjacent to open water, with the exception of marsh expansion at Mockhorn Island, a relatively protected site near the mainland. Erwin *et al.* (2006) determined that relative sea-level rise at Wachapreague exceeded the rates of accretion for the mid marsh and high marsh. With sea level rising at 3.9 mm/yr, they estimated that the elevation of the high marsh increased 1.4 ±0.2 mm/yr and the mid marsh 0.7 ±1.2 mm/yr. Marsh ponds, however, are filling at a substantial rate, 11.5 ±1.5 mm/yr, well above that of the relative sea-level rise. They observed that over the four years of their study, *Spartina* moved laterally into pond edges and tidal flats. However, they suggested that over the next 50 to 100 years, "lagoonal marshes may revert to open water." This contrasts with Finkelstein and Ferland's (1987) finding that the area of marsh in the back-barrier was expanding.

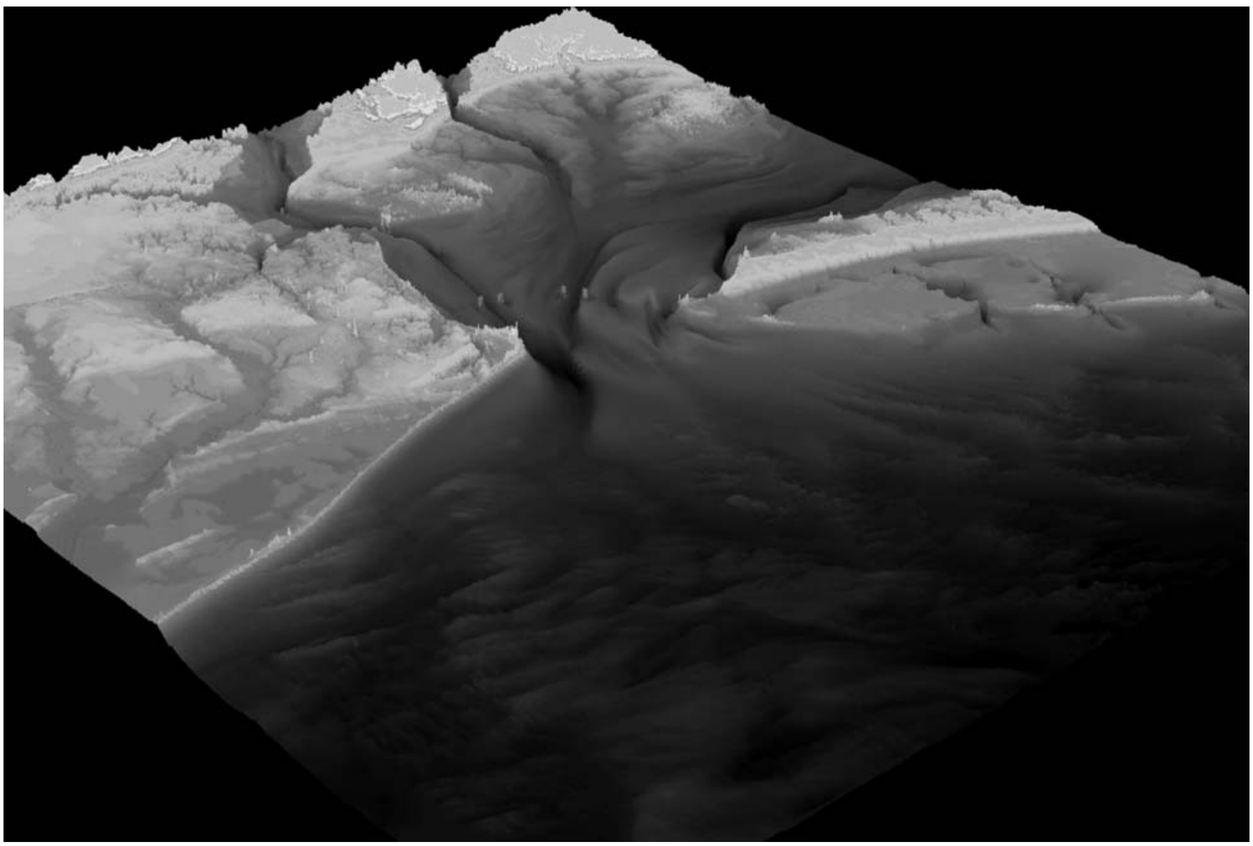


Figure 8: An oblique view depicting the topography of southeastern Virginia, the lower Delmarva Peninsula, the Chesapeake Bay mouth region, and adjacent portions of the continental shelf. The southwestern half of the Chesapeake Bay Impact Crater is expressed geomorphically by the arcuate escarpment starting at Cape Henry (at the southern margin of the Chesapeake Bay mouth) and continuing clockwise past the peninsulas separating the James, York, and Rappahannock Rivers.

The unmodified character of the tidal inlets of the Virginia Eastern Shore makes them inviting sites for studying inlet processes. Building on the efforts of O'Brien (1931, 1969), Byrne and co-workers (DeAlteris, 1973; Byrne *et al.*, 1974, 1977; Boon and Byrne, 1981) investigated the characteristics of Wachapreague Inlet. Byrne *et al.* (1974) concluded that the ratio of ebb tidal power to wave power is useful for characterizing short-term response of the inlet channel. They found that short-term modulations of the cross sectional area of the inlet commonly were due to sand moving between the tidal channel and the ebb-tidal delta complex. Li (2002), working in Sand Shoal Inlet, studied the water-mass fronts that develop in the inlet. He concluded that the dominant factor contributing to the generation of the fronts is "frictional tidal motion strongly affected by bathymetry and geometric constriction at the entrance."

Chesapeake Bay Mouth

The mouth of Chesapeake Bay is a complex and dynamic area that includes shoals and channels both inside the bay and on the inner continental shelf (Figure 8). Major shoals have developed on the interfluves between modern channels, the courses of which are influenced by ancient channels of the Susquehanna, York, and James Rivers. The Bay Mouth Segment includes the shoal-retreat massif of Swift (1975), who described this class of feature as "low, broad, shelf-transverse bodies which mark the retreat paths of coastal depocenters associated with littoral drift convergences."

Using sea-bed drifters, Harrison *et al.* (1964) documented bottom transport of sediment into Chesapeake Bay through the bay mouth. Meade (1969, 1972) discussed the net movement of sediment from the shelf into estuaries on the Atlantic coastal plain. Although he mainly considered suspended sediment, he indicated that there was evidence supporting the landward transport of sand as well.

Meisburger (1972) investigated the entrance to Chesapeake Bay for deposits of sand potentially usable for beach nourishment. This study was based on a suite of 57 ten-cm (four-inch) diameter vibracores up to 6 m (20 ft) in length obtained for planning the Chesapeake Bay Bridge Tunnel, along with many kilometers of seismic profiles collected for the project. Most of the surficial sediment in the bay mouth is fine to very fine sand that has been derived from outside the bay during the late Holocene. Coarse, gravelly sand from Pleistocene or older fluvial deposits is exposed in Thimble Shoals Channel, the southernmost channel into the bay.

Ludwick (1981), expanding on earlier works (Ludwick, 1970, 1975) described sediment trapping and accumulation of the bay mouth shoals. He found that the rates of deposition in the vicinity of Thimble Shoal Channel in the southern portion of the bay mouth were very high compared to literature values for other portions of Chesapeake Bay, and that the Tail of the Horseshoe, a shoal just north of the channel, was aggrading on its southern flank, in effect migrating south. He also noted that sand waves approximately 5 km (3 mi) north of the Tail of the Horseshoe were asymmetrical in shape, indicating movement to the south. Granat and Ludwick (1980) determined that Inner Middle Ground/Nine-Foot Shoal, which is 10 km (6 mi) northeast of Tail of the Horseshoe, had migrated 1.9 km (1.1 mi) southward in 124 years. They describe the shoal as being "in quasi-equilibrium with mutually evasive ebb and flood tidal currents." The relief of the shoal exposes one face of the shoal while protecting another from either flood or ebb currents. Ludwick (1970) described Inner Middle Ground/Nine-Foot Shoal in the northern portion of the bay mouth as having formed between ebb-dominated and flood-dominated channels.

Colman *et al.* (1988) stated that the interaction of processes related to the growth of the tip of the Delmarva Peninsula, tidal current reworking of sediments transported to the bay mouth, and net landward transportation of sediment are responsible for the configuration of the bay-mouth shoals. They indicated that the present shoal is a distinct stratigraphic unit whose "position and internal structure .... show(s) that it is related to near-present sea level, and thus is less than a few thousand years old." Their work further concludes that "the source of the bay-mouth sand is primarily outside the bay in the nearshore zone of the Delmarva Peninsula and on the inner continental shelf."

Hobbs *et al.* (1990, 1992) determined that a substantial quantity of sandy sediment moves from the shelf into Chesapeake Bay. They deduced that the continental shelf was by a very large margin the principle source for sediment deposited in the lower Bay. Seismic profiles presented by Colman *et al.* (1985) demonstrate substantial southward and westward growth of the bay mouth shoal into the

Bay. These studies are bolstered by the development of Fishermans Island at the southern tip of the Delmarva Peninsula. Boule (1976a, b, 1978) documented the growth of the island from shoals on late 18th century charts, to its establishment as an island by 1815, and continued growth through the time of his study.

Oertel and Overman (2004) expanded on the theme of Fishermans Island. They differentiated it from the more classical barrier islands of the ocean coast and characterized it as having formed as a consequence the convergence of the southerly paths of sediment transport on the ocean and bay sides of the Delmarva Peninsula. The resulting accumulation of sediment is further modified by wave-driven processes, primarily from the ocean side. Additionally, the refraction and diffraction of waves at the shoals and bay mouth serves to generate linear bars along shore that migrate landward. The ridges and hummocks of the modern geomorphology of the island records the history of this process.

Together, these studies indicate that the Bay Mouth Segment of the coast of Virginia has an important role in the sediment transportation and deposition processes of the inner and mid-continental shelf.

Swift *et al.* (2003) used the Quaternary sediments of the broad region of the Chesapeake Bay mouth as a case study in hierarchical stratigraphy. They considered the relationships between stratigraphic units of small spatial extent and short times of formation and the physically and temporally larger scale units employed in sequence stratigraphy. They viewed “mesoscale” or “event scale” units as, in part, being defined by grain-size characteristics and as having generally finer sediment downstream. They considered that after some thickness of the surface sediments are resuspended or reactivated by an event such as a storm, the coarser sediments are the first to be redeposited and are covered by finer materials. As less intense events occur more frequently than major events, the uppermost portion of the sediment column is reactivated more often, thus progressively moving finer sediments to the surface where they are more subject to downstream transport. Major storms that agitate the bottom to a much greater depth are relatively rare but tend to concentrate the coarser sediments at the bottoms of the lesser units above. The resulting strata occur within the larger spatial and temporal context of a rising, or falling, sea level and the resultant higher-order sequence of strata.

Southeastern Virginia Mainland

As noted previously, the shore of southeastern Virginia is part of the large coastal compartment that extends from Cape Henry at the mouth of Chesapeake Bay south to Cape Lookout. The Pleistocene mainland forms the shoreline in the northern portion of Virginia Beach. The surficial deposits of the mainland are of the upper Pleistocene Tabb Formation (Mixon *et al.*, 1989), and its Poquoson and Lynnhaven Members intersect the shore. A few kilometers inland, the Sedgefield Member roughly parallels the shore and has the gross morphology of a chain of barrier islands (Figure 8). The Pleistocene islands form the seaward edge of the lowlands now occupied by the Dismal Swamp which continue westward to other exposures of the Sedgefield Member along the Suffolk Scarp, a late Pleistocene mainland shoreline. Hobbs (2004) discussed the history of the regional stratigraphic nomenclature.

The shore of southeastern Virginia has been the site of many studies concerned with shoreline processes, especially as they might affect the important beaches. Harrison and Krumbein (1964) considered the coastal processes in the vicinity of Virginia Beach. They determined different combinations of process variables affected the shore zone in different ways and that in some instances, there was an 8 to 12 hour time lag between the process activity and the observable response.

Goldsmith *et al.* (1977) reported repetitive surveys of 18 beach profiles spaced between Cape Henry and the Virginia–North Carolina border during the period of September 1974 through December 1976, and reviewed profile data going back to 1956. They found narrow, erosional beaches in the center of the area, Dam Neck, Sandbridge, and Back Bay National Wildlife Refuge, and wider accretional beaches at the north and south ends of the study area, Ft. Story and False Cape State Park, respectively. During the study period, the maximum rate of beach accretion was 18.9 m^3 per meter of beach front per year at Fort Story, as compared to the maximum rate of erosion of 11.6 m^3 per meter per year at Sandbridge. They suggested that this pattern exists because the central area is a nodal zone in which the direction of net longshore transport shifts from north to south. They also stated, “The Virginia Beach commercial area would be erosional without the extensive sand nourishment which is necessary for the maintenance of the commercial beaches."

The southern portion of the shoreline of southeastern Virginia is part of Currituck Spit, the long barrier spit that extends from Sandbridge, Va. to Oregon Inlet, N.C., and constitutes the northern segment of the "Outer Banks." Hennigar (1977) described the history of the barrier island since earliest colonial times. Although there are numerous indications that inlets had existed in the past, only "Old Coratuck" (or "Old Curritucт") Inlet, which opened in 1657 and closed in 1728, has been known in historic times (Hennigar, 1977; Prow, 1977). Indeed, according to Sharpe (1961), Currituck was a major port. In 1728, while surveying the border between Virginia and North Carolina, William Byrd described the inlet as follows: "It was just Noon before we arrived at Coratuck Inlet, which is not so shallow that the Breakers fly over it with a horrible Sound, and at the same time afford a very wild Prospect. ... leaving an opening of not quite a mile, which at this day is not practical for any vessel whatsoever. And as shallow as it is now, it continues to fill up more and more, both the Wind and the Waves rolling in the Sands from the Eastern Shoals."

Recent work in the former Currituck Inlet (McBride *et al.*, 2004; Robinson and McBride, 2004, 2006) documented the dynamic nature of inlets and the Outer Banks barrier. The inlet was open before 1585. Between 1713 and the closure of Old Currituck in 1731, a new inlet, New Currituck Inlet, opened about 10 km south, and progressively captured the tidal prism of the back barrier. Through analysis of the lithology and x-radiographs of vibracores from the relict flood-tidal delta, McBride *et al.* (2004) identified four sedimentary facies that record the history of the inlet. From oldest to youngest, these sedimentary environments are the 1) estuary, 2) active, subtidal, flood-tidal delta, 3) diminishingly active, intertidal, flood-tidal delta, and 4) marsh. Studies of the foraminifera (Robinson and McBride, 2006) also show salinity in the marsh decreasing with time. The vibracores and ground-penetrating radar (GPR) images (McBride *et al.*, 2004) depict the paleochannel as about 400 m (1,300 ft) wide and up to 7 m (23 ft) deep, and as having prograded from north to south.

Much of the ocean shoreline of southeastern Virginia has been managed through construction of sea walls and/or nourishment. Three separate areas, the resort area north of Rudee Inlet, Sandbridge, and the federal facility at Dam Neck, have been nourished at various times during the past half century. Between 1951 and 2003, almost 17.5 x 10^6 m^3 (23 x 10^6 yd^3) of sand were placed during approximately 50 separate projects for a total cost of about $150 million (Western Carolina University, 2007). Most of the projects were annual renourishment efforts along 5,600 m (18,480 ft) of the resort area. Much of this material has been obtained from maintenance dredging of Rudee Inlet, occasional dredging of entrance navigation channels to Chesapeake Bay, and, early on, upland sources.

Since 1996, Sandbridge Shoal has been the source of material for nourishment at both Dam Neck and Sandbridge. Through 2003, just over 3.5 x 10^6 m^3 (4.6 x 10^6 yd^3) (Diaz *et al.*, 2006) of sand have been removed from the shoal and placed on the two beaches. Wikel *et al.* (2006) document the recent history of shoreline change of the area. Hardaway *et al.* (1998) reviewed many of the previous studies of shoreline change. In the vicinity of Rudee Inlet there has been some accretion that most probably is associated with sand trapping by the jetties. Also, the area immediately south of Rudee Inlet benefits from its position relative to the sediment transport nodal zone whereby, locally, sand is transported north into the area which also abuts late Pleistocene outcrops. Farther south, where the shoreline is backed by marsh or lagoon, there is a history of long-term erosion and shoreline retreat.

Williams (1987a,b) investigated the characteristics of sand bodies with sediment suitable for use in beach nourishment projects and to enhance the understanding of the Quaternary geology of the area. The study involved collection of 302 km (163 n mi) of seismic profiles and analysis of 138 previously obtained vibracores. The study basically affirmed the stratigraphic interpretation of Shideler *et al.* (1972) and affirmed their Unit D, a discontinuous, surficial sand sheet, that has formed during the Holocene.

The filled, near-surface channels described by Chen (1992) and Chen *et al.* (1995) may be the same set of channels later described by McNinch (2004), Browder (2005), and Browder and McNinch (2006), that influence both morphology of the nearshore and the surficial sediment type. McNinch (2004) studied a 16 km (10 mi) reach of the shorezone between Dam Neck and False Cape on the southern coast of Virginia and a larger section of the northern portion of the Outer Banks of North Carolina with a variety of geophysical, remote-sensing techniques. He noted that shore-oblique sand bars existed adjacent to gravel outcrops, and that those outcrops are surface expressions of underlying strata. After energetic conditions, large scale sandbars recurred in the same locations, and there is a spatial relationship between the sandbar-gravel outcrop pairs and erosional hotspots. Schupp *et al.* (2006) further developed the correlation among the shore-oblique bars, gravel outcrops, and shoreline change, although this study was confined to the northern Outer Banks.

Browder (2005) and Browder and McNinch (2006) linked the antecedent geology and the nearshore morphology in the Sandbridge, Virginia and the North Carolina study areas. They also mapped one broad and three relatively narrow, southwest-northeast oriented channels crossing the nearshore zone at Sandbridge. The smaller channels, which span about 100 m (330 ft) bank to bank, perhaps are analogous to the "Cape Charles Aged Paleochannels" of Chen (1992) and Chen *et al.* (1995) and the small channels depicted by Hobbs (1997). It is likely that these channels were excavated during the lowstand of the latest Pleistocene into the early Holocene, and were filled during the ensuing transgression. They could be either old upland drainage or tidal inlet channels. Though having the same orientation, the larger, roughly 800 m (2,600 ft) cross section, channel is more problematical as to origin. Hobbs (1997) noted three channels of similar scale but different orientation farther offshore. The oldest of these channels is aligned northwest-southeast, whereas the other two approach a north-south orientation and more closely parallel the modern shoreline.

Hardaway *et al.* (1995) identified fluvial sediments in several cores roughly 5 km (3 mi) north-northeast of Rudee Inlet. They obtained a ^{14}C date of 9,440 ± 50 yr from wood at the base of a fluvial unit 3.7 m (12 ft) below the sea floor, roughly 15 m (50 ft) below sea level. As Bratton and others (2003) indicated, sea level in the Chesapeake Bay region did not reach this level for another thousand years or so. Consequently, these channels most likely were carved by upland drainage.

EVOLUTION OF THE VIRGINIA COAST

As discussed in the introduction to this chapter, the modern coast of Virginia is being shaped by wave and tidal energy while undergoing a long-term rise in sea level since the last glacial lowstand 20 to 22 ky ago. The present coast also has a geologic legacy, or foundation, inherited from the sedimentary processes operating during the preceding sea-level highstands and lowstands driven by the global glacial-interglacial cycles of the Pleistocene.

Relative Sea Level

Change in sea level, more properly in relative sea level, is a major factor in the evolution of the Mid-Atlantic coast of North America. *Relative sea level* is the vertical position of the sea relative to fixed landmarks; it can be thought of as the effective sea level. Two major sets of processes work together to change relative sea level: eustatic processes and regional or local tectonic processes. Eustatic processes influence the level of the sea globally. Factors that change the volume of water in the world ocean or the volume of the world-ocean basin are eustatic processes. Variation in the amount of water trapped in the world's ice caps is the most well known eustatic process. The polar ice sheets wax and wane in cycles of approximately 20 and 40 ky, and in major cycles of 100 ky that produce sea-level changes on the order of 120 m (395 ft). Other processes, such as the addition of new or "juvenile" water from subsea volcanoes, also add water to the ocean but at a much slower rate. Physical processes associated with sea floor spreading alter the size of the ocean basins and thus the level of the sea. Similarly, sedimentation, the transfer of sediment from the continents to the ocean basins, partially fills the ocean basins thus displacing sea level upward. The last two sets of processes work much more slowly than the freezing and thawing of the ice caps.

Changes in regional to global climate and extent of glacial ice can be reconstructed from many different preserved geologic records, but some of the most continuous and detailed records are from cores drilled in glacial ice and deep-ocean sediments. Many of the ice cores from the Antarctic and Greenland ice sheets have annual layers of snow that provide a chronology (timescale) analogous to that of tree rings. These cores yield very detailed records of changes in temperature and atmospheric gases, including the greenhouse gases carbon dioxide and methane, back through the most recent ice age (roughly 10 to 60 ka, or thousand years ago) to the previous interglacial warm period (about 80 to 130 ka). Recent work with extremely long ice cores has yielded records through multiple glacial-interglacial cycles as far back as 400 to 800 ka (Jouzel *et al*., 2002, 2007). Changes in temperature over the ice sheet are quantified by analyzing the ratios of stable (i.e., not radioactive) isotopes of oxygen (^{16}O and ^{18}O) and hydrogen (^{1}H and ^{2}H) in the water molecules of the ice.

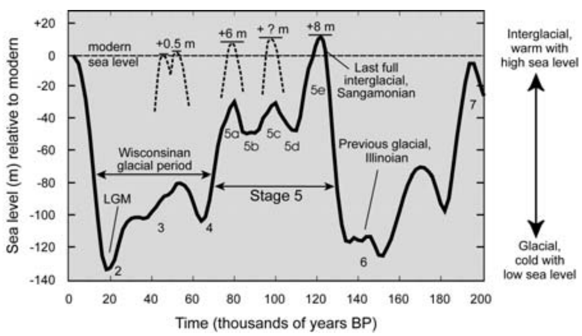


Figure 9: Oxygen isotope record of global ice volume and sea level for the past 200,000 years showing the major glacial and interglacial episodes (data from Martinson *et al*., 1987). Dashed lines above the curve indicate highstand deposits in the Mid-Atlantic region associated with the global periods of reduced ice volume (Wehmiller *et al*., 2004; Mallinson *et al*., 2008).

The climate records derived from sediment cores from the deep ocean also rely on stable oxygen isotopes, but as preserved in the calcium carbonate shells of microscopic marine organisms such as foraminifera. These tiny protists live both up in the water column (planktonic) and in the mud of the sea floor (benthic). Similar to the geochemistry of ice, the ratio of ^{16}O to ^{18}O in the $CaCO_3$ of the shells records the water temperature when and where the foraminifera were living. However, the isotopic composition of the shell also records the isotopic composition of the seawater itself. As water evaporates from the surface of the ocean and is transported through the atmosphere toward the poles, there is a net removal of water molecules with the heavier ^{18}O by the process of fractionation. If enough water carrying ^{16}O is removed from the ocean and stored as glacial ice, the water remaining in the ocean is enriched with ^{18}O; the degree of enrichment is proportional to the global ice volume. Consequently, isotope records from the planktonic foraminifera living near the ocean surface have a mixed climatic signal that is part temperature and part ice volume, whereas the benthic records, from deep ocean basins where water temperature is constantly cold, are an excellent indicator of global ice volume.

Cyclic variations in the oxygen isotope ratios occur throughout the world ocean and ages have been determined for a long period of geologic time (Hays *et al*., 1976; Lisiecki and Raymo, 2005). The cycles have been enumerated with the convention that odd-numbered marine isotope stages indicate interglacial (warm) intervals and even-numbered stages indicate glacial (cool) stages (Figure 9). Marine Isotope Stage 2 was the most recent glacial episode – coinciding with the lowstand of sea level that reached a minimum at approximately 21 ka. Stage 3 was only a mild warming and global sea level did not reach the present level, and Stage 4 was a mild cooling that was not as extreme as the glacial maximum of Stage 2. During Stage 5, roughly 73 to 127 ka, sea level reached and slightly exceeded the present level. Stage 5 sea level fluctuated with three peaks and intervening lows that are designated substages 5a through 5e (Figure 9). Toscano (1992) and Toscano and York (1992) presented a sea-level curve covering the past 130 ky for the Mid-Atlantic.

Tectonic processes, often called isostatic processes, result in regional or local movements of portions of the Earth's crust relative to adjacent portions. If an area is uplifted relative to the contemporaneous elevation of the ocean, the local sea level appears to fall; conversely, if an area subsides, sea level appears to rise. These processes may act rapidly due to major earthquakes, but more typically they act slowly, such as the progressive depression of the land surface under and near growing ice caps, or as rising sea level floods the continental margin. Because the Earth's mantle flows slowly, the response of vertical movement often lags the change in loading at the surface, and may continue to adjust for thousands of years.

Change in relative sea level is the sum of the eustatic and isostatic processes. The change in relative sea level can vary from place to place because although the eustatic component is uniform around the world ocean, the isostatic-tectonic component can differ between regions. Relative vertical motion can be dramatically different within a small distance, such as in Alaska where two sites can be on opposite sides of an active fault.

Beyond the global to regional effects of eustatic sea level and tectonics, local factors controlling the response of a section of the coast include the topography, antecedent geology, and sediment supply. Even during an overall rise in sea level, different sections of a coast may experience different rates of rise or even a relative fall of sea level depending upon the interplay of these factors. For example, if eustatic sea level is rising, the land is subsiding, and/or sediment supply is relatively low, the shoreline or barrier islands will migrate landward, producing a *transgression*. Conversely, if eustatic sea level is falling, the land is being uplifted, or sediment input to the shoreface is substantial, the shoreline migrates seaward, producing a *regression*. Sediment supply in particular may vary considerably along a stretch of coast – a spit that is the receiving end of longshore transport may be prograding (extending) while at the same time a sediment-starved downdrift area may be eroding and retreating rapidly landward. This is the case with the progradation of southern Assateague Island and the rapid retreat of the downdrift barrier islands of the Chincoteague Bight.

In the Mid-Atlantic region, subtle tectonic processes add to the eustatic component with the result that the rate of sea-level rise as determined from long-term tide gages generally is between 3.0 and 4.4 mm per year (1 to 1.5 feet per century) (Zervas, 2001). The greatest present rate of rise in the area is measured at Hampton Roads, within the faulted and subsiding Chesapeake Bay Impact Crater. Because the Mid-Atlantic continental margin has such a low slope, generally less than one degree, 1 m (3.3 ft) of sea-level rise typically results in 50 to 120 m (165 to 395 ft) of landward migration of the shoreline (Zhang *et al*., 2004). As sea level rises and the high-energy shoreline moves landward, any low-lying areas behind the shoreline are progressively inundated to create the back-barrier lagoon and marsh complex, and river valleys are flooded to produce estuaries with the influence of tides extending 10s to 100s of kilometers landward of the ocean coast.

Holocene Sea-Level Rise

As noted above, the most dramatic eustatic changes in sea level result from the freezing and thawing of the polar ice caps, primarily in the Northern Hemisphere. Since the last glacial maximum at about 21 ka, sea level has risen close to 125 m (410 ft) (Peltier, 2002). For much of the twentieth century, it was assumed that sea level rose quickly but steadily from its minimum until about 5 ka, then slowed significantly (for example, Milliman and Emery, 1968). More recent work (Fairbanks, 1989; Nikitina *et al.*, 2000; Peltier, 2002) indicates that sea level has risen since the last glacial maximum in a series of steep steps with intervening plateaus that might have included minor regressions. The periods of exceptionally rapid sea-level rise appear to have been in response to major pulses, or releases, of meltwater from the Northern Hemisphere ice sheets. During the most prominent of these meltwater pulses at about 12 and 9.5 ka, sea level was rising 10 to 20 times faster than at present, and transgression of the shoreline across the coastal plain would have been clearly evident within a few decades. As sea level rose above about 18 m below the present level, between 8.2 and 7.2 ka, brackish waters reached northern Chesapeake Bay (Bratton *et al.*, 2003).

The rate of sea-level rise slowed considerably by about 6 ka (Bratton *et al.*, 2003), when the last vestiges of the major ice sheets covering Canada and Scandinavia had melted completely. There is substantial evidence from low-latitude sites for a mid-Holocene highstand, estimated between 5 and 1.5 ka, with relative sea level between 1 and 3 m higher than present (Grossman *et al.*, 1998; Morton *et al.*, 2000; Blum *et al.*, 2001; Peltier, 2002). In quantitatively modeling Holocene sea level, Peltier (2002, 2004) considered the slow, complex, viscous flow within the Earth's mantle in response to changes in load on the crust from melting ice sheets and the increasing volume of ocean water, and the time lags between the change in load and response of the crust.

Some of the earliest work on Holocene sea level along the Virginia coast was that of Harrison *et al.* (1965), who evaluated cores and seismic lines collected for construction of the Chesapeake Bay Bridge-Tunnel. They interpreted what appeared to be anomalously shallow depths of an assumed buried Susquehanna River channel, and the occurrence of intertidal peats and underlying shell beds with ^{14}C dates from 1,170 to 1,900 years BP as much as 1.5 m (5 ft) above sea level. From these observations, Harrison *et al.* (1965) postulated a regional uplift in the southern Chesapeake Bay during the late Pleistocene or Holocene. This hypothesis was supported by Newman and Rusnak (1965), who obtained radiocarbon dates from buried tidal-marsh peats in the vicinity of Wachapreague. The maximum thickness of Holocene sediments in the area was approximately 11 m (36 ft), and the oldest date they obtained was 5,120 ± 145 years from a sample 6.1 m below the marsh surface. Another sample, almost 800 years younger (4,350 ± 75 years) was obtained from a depth of about 7.2 m, roughly a meter shallower than the older sample.

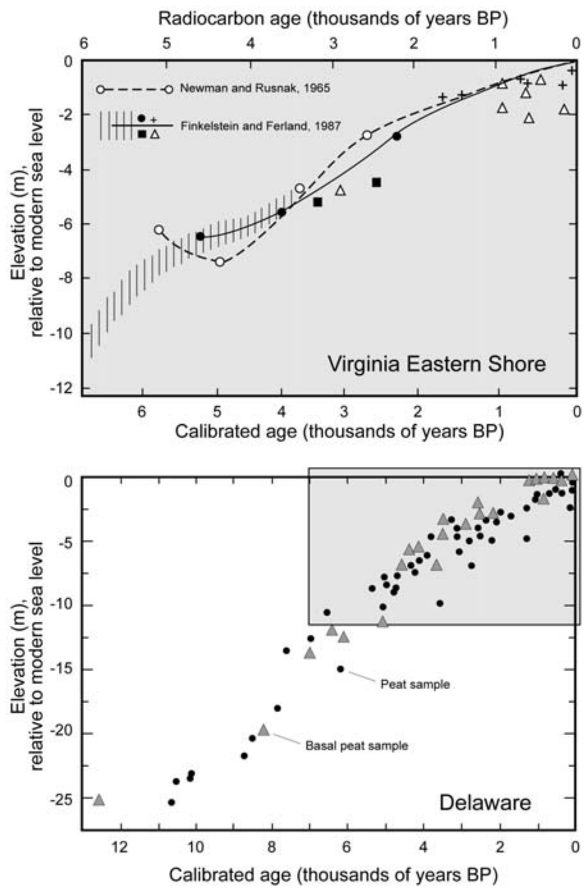


Figure 10. Holocene sea-level curve for the Virginia Eastern Shore (upper panel), modified from Van de Plassche (1990). A more extensive, regional sea-level curve for the Delaware coast (lower panel), mainly from the work of Kraft and his students, as compiled by Ramsey and Baxter (1996). The shaded inset in the lower panel is the equivalent age and elevation range as the upper panel.

Newman and Munsart (1968) continued to work near Wachapreague, and determined that the barrier islands and the back-barrier lagoons have existed for at least 5,500 years. They also suggested that Italian Ridge, a prominent dune ridge oriented about 20º clockwise of the regional shoreline on Parramore Island, was formed during the Holocene and "may have been constructed along a prograding shore, perhaps during a period of emergence." Coupling the high sea level suggested by Italian Ridge with a several-meter fall of sea level between 5,100 and 4,400 years BP, they inferred either a brief regression or a "cessation of crustal uplift." Either of these situations require that sea level was a meter or more above the present level several thousand years ago. This view was supported by Kemerer (1972), who concluded that the presence of features such as Italian Ridge indicates that there was a "short term regressive phase during the migration of (the barrier) islands."

Harrison (1972) studied the intertidal areas of the Atlantic coast of southern Delmarva and characterized the biota, sediments, and sedimentary processes of the back-barrier mudflats and marshes. His work specifically "failed to reveal any elevated peats that would support the concept of local uplift during the late Holocene." He also noted the existence of several other ridges oriented similar to Italian Ridge, and suggested that they are possibly Pleistocene features rather than Holocene.

Finkelstein and Ferland (1987) related the relative rise of sea level and/or a local deficiency in sediment to the Holocene retreat of the barrier islands of the Virginia Eastern Shore. They attributed the diminution of the back-barrier area, a resulting decrease in tidal prism that yields a decrease in the cross-sectional area of the tidal inlets, and accelerated rates of accretion in the marshes and tidal flats to those driving factors. Their sea-level curve for the past 5000 years shows a slightly increased rate of rise between about 3.8 and 2.2 ka, when sea level rose from about -5.5 to -2.7 m (Figure 10). They stated that the present rate of rise was between 2.0 and 3.6 mm/yr, or 0.5 to 2.1 mm/yr faster than for the period beginning at 4.6 ka. Their analysis of cores taken in the back-barrier region indicated that "extensive marsh growth occurred after approximately 1.6 ka," when the rate of sea-level rise slowed. They also stated that new marsh growth continues on the tidal flats, washover fans, and flood-tidal deltas.

Van de Plassche (1990) evaluated the past 6000 years of sea-level change on the Virginia Eastern Shore by reviewing the previous studies and adding new radiocarbon dates (Figure 10). He concluded that "the trend of the [relative sea-level] rise before 4000 yrs B.P. was as rapid and continuous as in all other states further north," including the extensive set of dates from the Delaware coast (e.g., Belknap and Kraft, 1977; recently compiled by Ramsey and

Baxter, 1996). He discounted the drop in sea level between 5.1 and 4.4 ka reported by Newman and Munsart (1968).

Antecedent Geology and Topography

Antecedent geology and topography influence both the modern morphology of the ocean coast and its response to sea-level rise. For example, most of the islands of the Virginia Barriers are anchored on the interfluves (ridges between streams) of the lowstand drainage network, whereas most of the tidal inlets and back-barrier tidal creeks align with the lowstand valleys (Oertel *et al.*, 1992; Oertel and Foyle, 1995). In this setting, the updrift, main body of the barrier island sitting on the interfluve is relatively stable, whereas the inlets and the downdrift spit end of the islands have higher and more variable erosion rates. Similarly, where Pleistocene or earlier Holocene shorelines intersect the modern ocean coast, the topographic high of the older ridge and the abundant source of sand that can be mined by wave action and redistributed into the shoreface will stall the transgression at least locally (Riggs *et al.*, 1995). These effects have dramatically slowed shoreline retreat at five sites along the wave-dominated segment of the Delmarva coast: central Assateague Island, Fenwick headland, Bethany headland, Rehoboth headland, and sections of Cape Henlopen (Honeycutt, 2003; Honeycutt and Krantz, 2003).

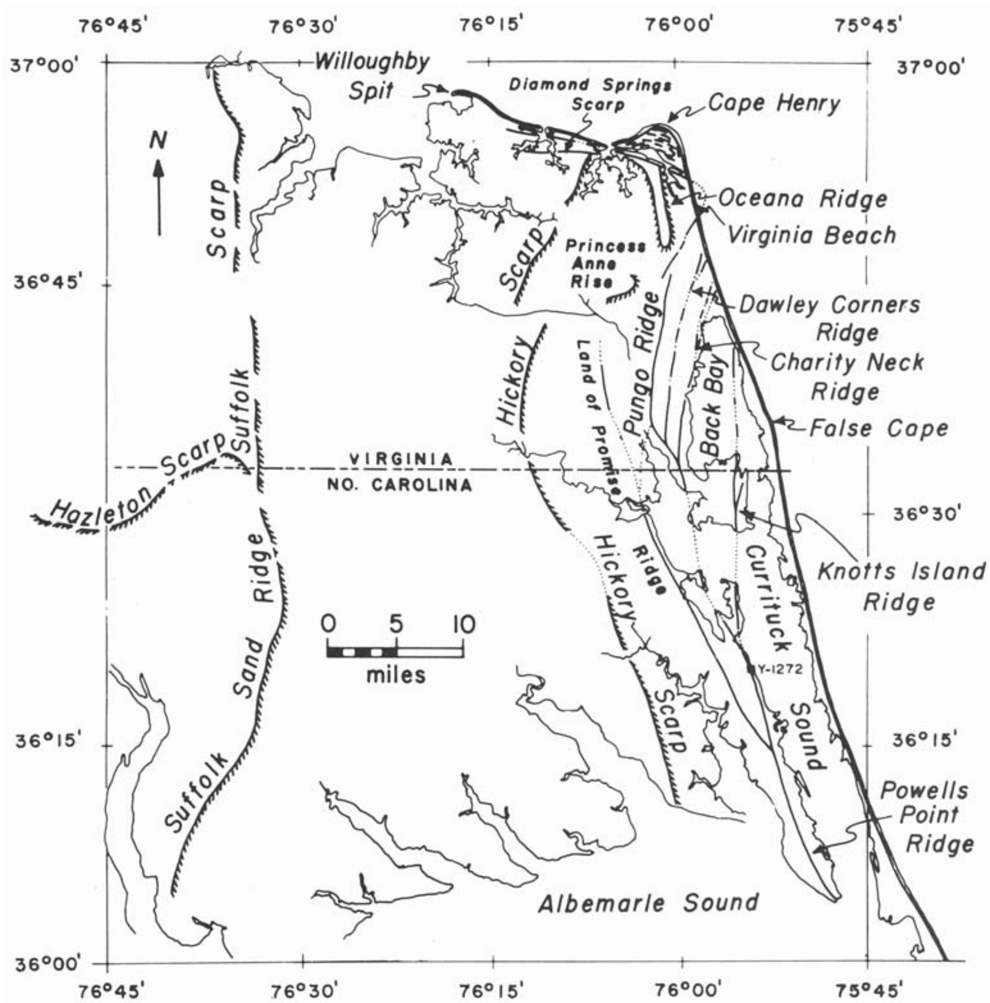


Figure 11. Late Quaternary scarps and shoreline ridges in southeastern Virginia and northeastern North Carolina, from Oaks *et al.* (1974).

The perspective view of the Virginia coast presented in Figure 8 illustrates the abrupt increase in slope of the upper shoreface within about 1 km (0.6 mi) of the beach. Erosion by storm waves and shore-parallel currents are effectively carving away the shoreline and redistributing the sediment: across the island as overwash, downdrift with littoral transport, and offshore obliquely into deeper water. As sea-level rise proceeds, the net erosion to water depths of 11-13 m (36-43 ft) (Kraft, 1971) produces the *shoreface ravinement surface*, which is a prominent marker horizon in high-resolution sequence stratigraphy (Swift *et al.*, 2003). As sea level falls, the highstand shoreline is stranded and the newly exposed coastal plain preserves a seaward terrace that previously was the shoreface, the scarp associated with the beach, the slightly elevated body of the barrier island, and a second, landward, terrace that was the back-barrier lagoon. The previous tidal inlets and tidal channels become the primary channels of the nascent drainage network developing on the land surface.

Looking landward across the lower coastal plain of southeastern Virginia on Figure 8, the land surface west of the modern shoreline is a series of flats separated by a few low-relief ridges, and cut by troughs or valleys presently occupied by bays or tidal tributaries. These terraces and shorelines were first described and named by Oaks and Coch (1973) and Oaks *et al.* (1974) (Figure 11), and the surface and subsurface units subsequently were revised and organized into the three members of the upper Pleistocene Tabb Formation by Johnson (1972, 1976) and Peebles (1984), as presented on the coastal plain map by Mixon *et al.* (1989). Visible on Figure 8, proceeding west from the modern shoreline are the Pungo ridges, Oceana ridge, Land of Promise ridge, Hickory ridge, and in the distance on the edge of the figure, the Suffolk Scarp cutting into the next-higher terrace on the mainland. Geomorphically and stratigraphically, these ridges have the appropriate characteristics for high-energy, open-ocean barrier islands.

The ages of these shoreline features have been debated for the past 40+ years, but they are generally considered late Pleistocene, and presumably represent stepwise regression at the end of the last interglacial episode. The Hickory ridge sits between 6 and 7.5 m (20-25 ft) above modern sea level, making it a good candidate for the Sangamonian or substage 5e highstand at about 127-120 ka, based on a prominent highstand worldwide at +7 to +8 m at that time (Figure 9). However, early studies employing uranium-series radiometric dating of coral fragments from barrier-island and back-barrier strata exposed in sand quarries (Womack, Moyock, and Gomez pits) cut into the Hickory ridge yielded ages of 70 to 85 ka (Cronin *et al.*, 1981; Szabo, 1985). Wehmiller *et al.* (2004) confirmed these younger-than-expected ages using a more advanced analytical method for U-series dating. These dates make the prominent Hickory shoreline in southeastern Virginia late Stage 5 (substage 5a), which was followed shortly by a moderate cooling and drop of sea level globally by 50-60 m (Figure 9).

Based on marine isotope records of global ice volume confirmed independently by U-series dates of coral terraces on Pacific islands, New Guinea and Barbados, global sea level during substage 5a was approximately 20 m below present (Lambeck and Chappell, 2001; Potter and Lambeck, 2003). Three questions arise from the finding that the Hickory shoreline has a substage 5a age: (1) Why is the substage 5a shoreline 26 m higher than expected along the Mid-Atlantic coast? (2) What age are the younger shorelines seaward of the Hickory ridge? and (3) Where is the substage 5e shoreline that is so prominent worldwide?

Effects of Isostatic Adjustment in the Mid-Atlantic

A series of recent papers from the coastal geology group at East Carolina University has dramatically revised our collective view of the late Quaternary history of the Mid-Atlantic coast. Mallinson *et al.* (2005) and Parham *et al.* (2007) demonstrated that multiple upper Quaternary transgressive-regressive cycles are preserved as sedimentary sequences beneath Albemarle Sound and its tidal tributaries, as interpreted from high-resolution seismic surveys supplemented with cores. Then Mallinson *et al.* (2008) reported optically stimulated luminescence (OSL) dates for several of the shorelines on the lower coastal plain of North Carolina that are either connected to or correlative with shorelines in southeastern Virginia. The OSL method obtains dates directly from the sand of the ridges, and is independent of the more standard carbon-14 or uranium-series radiometric dates (Szabo, 1985; Muhs *et al*., 2002) or geochemical methods such as amino-acid racemization (Wehmiller *et al*., 1988).

Included in the findings of Mallinson *et al.* (2008) are documentation of:

- An initial late Holocene transgression dated between 3 and 2.5 ka, preserved as a series of beach ridges behind the modern beach at Kitty Hawk
- Stage 3 ages for Powells Point and Land of Promise ridges
- Stage 3 ages for the seaward side of Hickory ridge
- Late Stage 5 ages for the seaward side of the Suffolk shoreline

The significance of these results is that the outer coastal plain and shelf of the Mid-Atlantic region has a very different preserved history of the sea-level oscillations of the late Pleistocene than other regions globally. Based on the marine isotope record as a proxy indicator of global ice volume and sea level, as supported by uranium-series dates on late Pleistocene coral terraces in tropical and subtropical areas, the highstand shorelines along the Mid-Atlantic coast for late stage 5 and stage 3 are anomalously high.

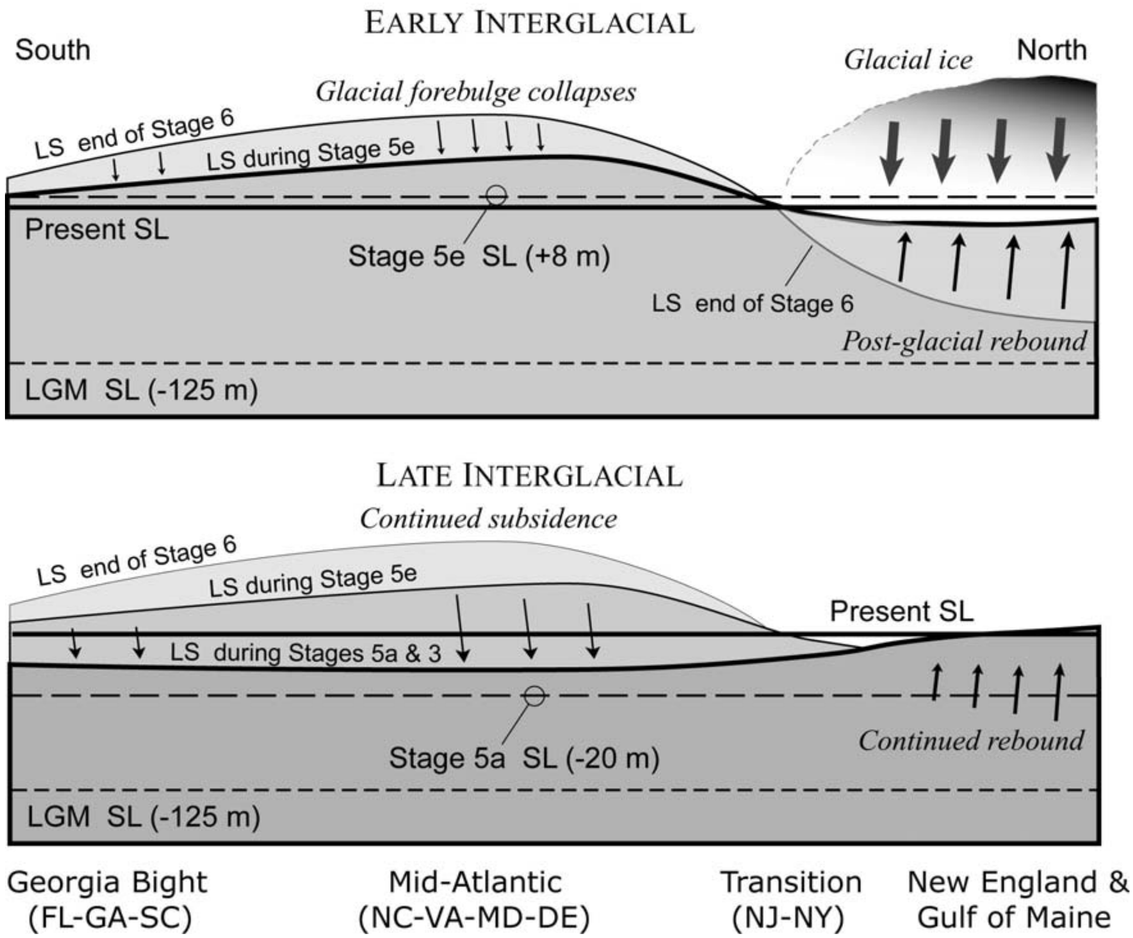


Figure 12. Regional isostatic response of the Atlantic coast after the Illinoian glaciation (marine isotope stage 6). This representation, based on the glacial isostatic adjustment model of Peltier (2004), follows from the recent findings by Wehmiller *et al*. (2004) and Mallinson *et al*. (2008) of late stage 5 and stage 3 highstand shorelines preserved on the outer coastal plain of Virginia and North Carolina approximately 22 to 26 m higher than expected from records of global ice volume. During the early part of the interglacial period following melting of the ice sheet, the glacial forebulge collapsed and retreated northward. The elevation of the land surface at that time (Sangamonian, or substage 5e) would have been roughly equivalent to the present. However, toward the later part of the interglacial, with continued subsidence of the forebulge in conjunction with hydroisostatic depression of the continental margin, the land surface was lowered by an additional ~25 m. [Abbreviations used in the figure: LS, land surface; SL, sea level; LGM, Last Glacial Maximum.]

Worldwide, the highstand of substage 5e was the highest sea level, and the warmest climate, of the past 200 ky (Figure 9), producing shorelines generally 7 to 8 m (23 to 26 ft) higher than present. Although the substage 5e shoreline has not been identified unequivocally in Virginia, the likely candidates are the higher-elevation part of the Sedgefield Member of the Tabb Formation (Johnson, 1976; Peebles, 1984; Mixon *et al.*, 1989) against the Suffolk Scarp (Figure 11; Oaks and Coch, 1973), and the prograding spit of the Butlers Bluff Member of the Nassawadox Formation on the Virginia Eastern Shore (Mixon, 1985; Mixon *et al.*, 1989). Both of these units have deposits in the elevation range of +8 to +10 m (26 to 33 ft), which is generally consistent with the substage 5e highstand globally.

The picture that emerges from these apparently anomalous elevations for highstand deposits for late Stage 5 and Stage 3 in the Mid-Atlantic is that the entire region underwent considerable isostatic adjustment after the Stage 6 glacial event (~140-160 ka) (Peltier, 2002, 2004). As discussed previously, when uneven forces are applied to the crust, by loading or unloading of mass, the upper mantle (specifically the partially molten asthenosphere) responds by flowing, slowly.

An appropriate physical model is pushing down with your thumb on a closed tube of toothpaste. As force is applied, the toothpaste, and the mantle, flow away from the downward force, creating bulges on either side of the depression. Along the east coast of North America, when glacial ice covered most of eastern Canada, New England, and the Gulf of Maine, the Mid-Atlantic sat astride the crest of the glacial forebulge (Figure 12). As the ice sheet melted and retreated to the northwest, the crust in New England rebounded and the forebulge in the Mid-Atlantic relaxed and migrated northward. Meanwhile, with the glacial meltwater flooding back into the ocean, sea level began rising, slowly at first and then rapidly.

The eustatic rise of sea level from the glacial maximum of Stage 6 to the peak highstand of substage 5e proceeded as rapidly as the equivalent transition from the Last Glacial Maximum of Stage 2 up to the present interglacial highstand. As sea level during substage 5e peaked and cut a shoreline notch into the coastal plain, much of the previous elevation of the forebulge had relaxed and, measured from the center of the Earth, the land surface in Virginia was approximately at the same position as today. However, through the remaining 40,000 years of the Stage 5 warm interval, the isostatic subsidence continued, although at a slower rate. In addition, the mass of the seawater that flooded the continental margin also was pressing down on the continental shelf and outer coastal plain. The result for the Mid-Atlantic margin was a total depression of at least an additional 22 to 26 m (Mallinson *et al*., 2008). This relative lowering of the land surface of the outer coastal plain allowed the late Stage 5 and Stage 3 highstands to flood the Mid-Atlantic almost to the same elevation as the peak substage 5e highstand, even though eustatic sea level was lower by 20 to 30 m (Figure 12). The implication of these findings and interpretations is that sea level along the Mid-Atlantic coast will continue to rise relative to the land well into the future, even without any further melting of glacial ice. However, most recent observations indicate that the rate of melting of montane glaciers worldwide, and sections of the Greenland and Antarctic ice sheets has accelerated dramatically in the last decade, resulting in an increasing rate of sea-level rise (IPCC, 2007).

The Middle to Late Pleistocene

As discussed throughout this chapter, the numerous oscillations of sea level during the Pleistocene and late Pliocene have formed the geology and shaped the geography of the Mid-Atlantic margin. The scarp and terrace topography of the coastal plain is a direct consequence of the cycles of transgression and regression, combined with the net long-term uplift of the Piedmont and coastal plain. Similarly the episodic lengthening of the Delmarva Peninsula (Mixon, 1985; Foyle and Oertel, 1997; Hobbs, 2004) occurred during separate highstands of sea level. Excavation of river channels beneath the modern Chesapeake Bay (Colman and Hobbs, 1987; Colman and Mixon, 1988; Colman *et al.*, 1990) and continental shelf occurred during times of low sea level when the ocean shoreline was at the modern shelf edge. Also during sea-level lows, the rivers transported sediment from the uplands to the shelf edge and onto the slope. The submarine canyons of the Middle Atlantic Bight likely were most active during these lowstands, transporting terrestrial sediments from the shelf into the deep ocean basin and aggrading the continental rise. Channel cutting and sediment transport would have would have been most vigorous during the initial phases of deglaciation as tremendous volumes of meltwater and glacial outwash sediments were channeled down the Hudson, Delaware, and Susquehanna River systems.

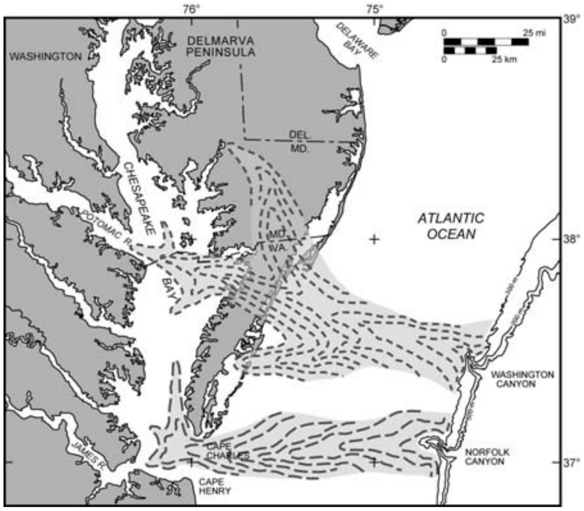


Figure 13. Inference by Harrison (1972) of a paleovalley of the Potomac River crossing beneath the northern section of the Virginia Eastern Shore, and heading to Washington Canyon. His interpretation included the recently discovered Salisbury (Maryland) paleochannel as a tributary to the Potomac River, and also connected the ancestral James River with Norfolk Canyon.

In the presently known history of the Virginia coast, four main channels of the combined Susquehanna-Potomac River system have been mapped under the Chesapeake Bay, the southern half of the Virginia Eastern Shore, and the inner shelf. However, Harrison (1972) speculated that a paleovalley of the ancestral Potomac River may have extended beneath the northern section of the peninsula (Figure 13) and pre-dated the Chesapeake Bay in its current configuration. His inference was based in part on the alignment of the mouth of the Potomac River estuary with Washington Canyon at the edge of the shelf. Additional supporting observations included the occurrence of coarse gravel and cobbles up to 25 cm diameter found in pits on the mainland west of Cedar and Wallops Islands. He also noted that gravel is only found on the beaches of the islands in the "peculiar indentation of the southern Delmarva shoreline of Wallops-Cedar islands," that is, the Chincoteague Bight. The lithology of the cobbles and gravels had a strong affinity to the Blue Ridge or Ridge and Valley provinces in the Potomac drainage basin. Harrison considered the recently discovered Naylors Mill (or Salisbury, Maryland) paleochannel (Hansen, 1966) to be a tributary to the main Potomac valley, although no evidence directly supports a correlation in space and time between the two.

If each of the four known paleochannels of the Susquehanna-Potomac system – the Exmore, Belle Haven, Eastville, and Cape Charles channels – is correlated with the maximum glacial events of the middle to late Pleistocene, the extension of the southern half of the Virginia Eastern Shore peninsula started about 350 ky ago (Figure 14). Prior to that time, each of the main rivers draining the Virginia coastal plain and Piedmont – the Potomac, Rappahannock, York, and James – probably flowed separately and generally southeast across the continental margin (Oertel and Foyle, 1995; Hobbs, 2004). However, their courses were almost certainly influenced by the depression (referred to as the Chesapeake Basin by Oertel and Foyle [1995]) caused by the Chesapeake Bay Impact Crater. Further, these pre-existing valleys were the topographic lows most likely to capture the Susquehanna-Potomac drainage as it was diverted southward.

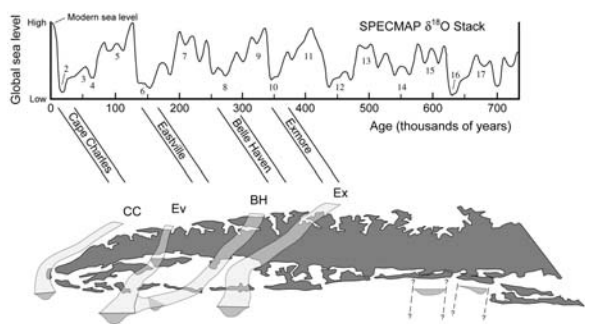
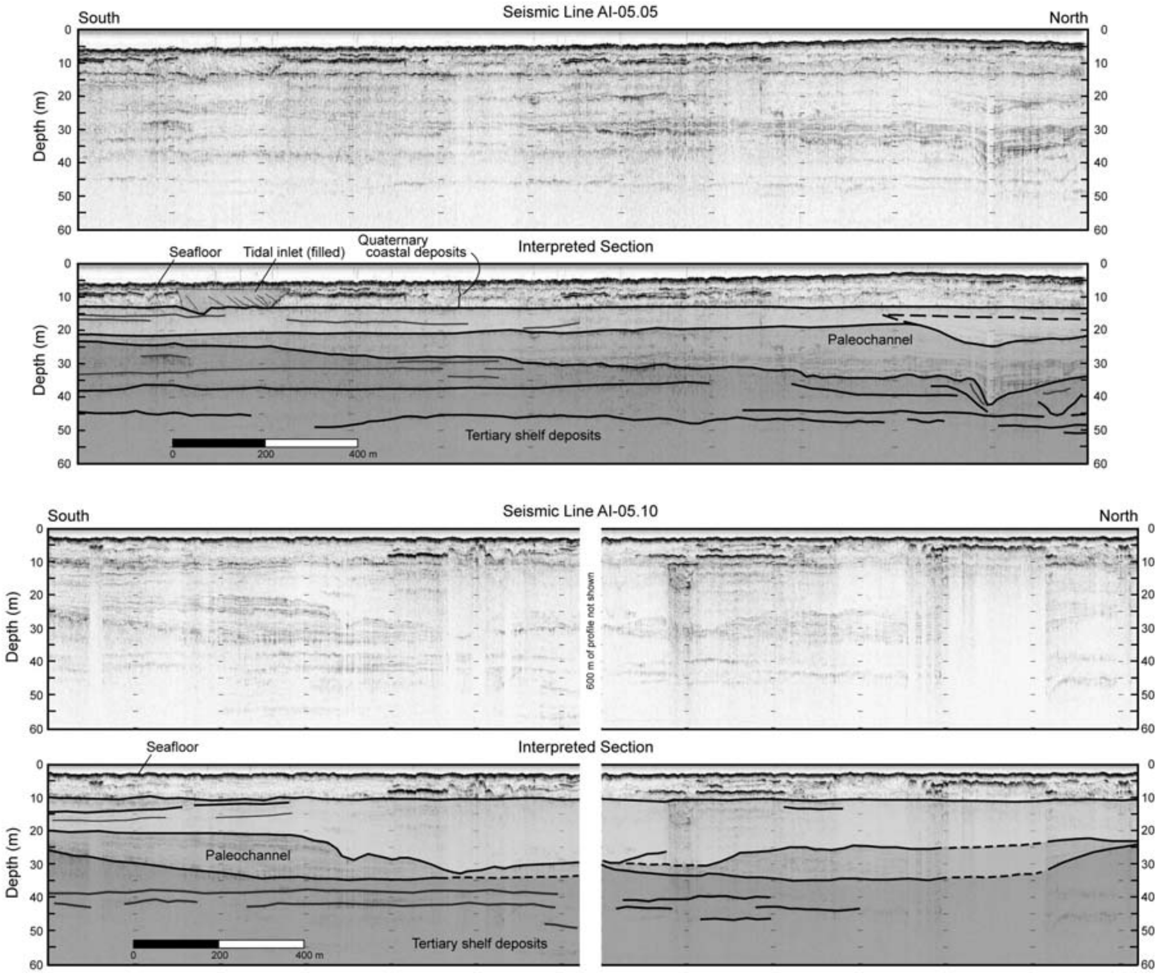


Figure 14. Paths of the four known middle to late Pleistocene paleochannels of the combined Susquehanna-Potomac River system crossing beneath the modern Virginia Eastern Shore (Colman *et al.*, 1990; Oertel and Foyle, 1995). The ages of these channels have been debated, but a simple chronology correlates each with the four full glaciations of the past 400 ky. At least two additional large paleochannels, possibly of an ancestral Potomac River, were discovered recently off Wallops and Assawoman Islands (Krantz *et al*., unpublished data).

Recent seismic surveys in the Chincoteague Bight, south of Assateague Island, found at least two large paleochannels off Wallops and Assawoman Islands (Figure 15) and extensive channelization into Tertiary strata below the apparent Quaternary coastal and shelf deposits (Krantz *et al.*, unpublished data). It is likely that these channels were produced by the Potomac and/or a combined Susquehanna-Potomac system. At a minimum, these features are middle Pleistocene, older than about 400 ky. There also is geomorphic and stratigraphic evidence for an earlier sequence of large-scale spit progradation, similar to the extension of the southern half of the peninsula, that was associated with closing off the Potomac River valley and rerouting the drainage to the south.



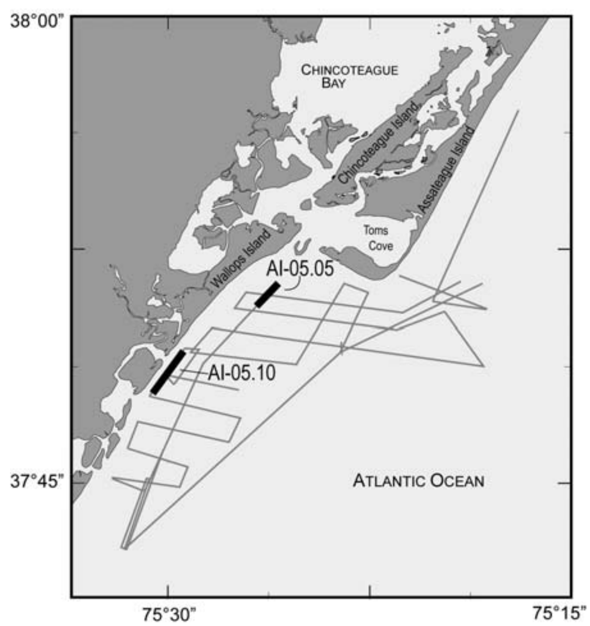


Figure 15a. Seismic sections of two paleochannels off Wallops Island (section AI-05.05) and off Assawoman Island (section AI-05.10) in the Chincoteague Bight with interpretation of primary stratigraphic units (Krantz *et al.*, unpublished data).

Figure 15b. Map of the northern Chincoteague Bight showing tracklines and seismic sections presented in Figure 15a.

FUTURE WORK

Studies Concerning Changes That Accompany Sea-level Rise and Climate Change

Sea-level change is a major factor in coastal studies that consider phenomena with time scales of a decade or more. NOAA (2007) states that from 1927 through 1999, the average rate of sea-level rise at the Sewalls Point tide station in Hampton Roads was 4.42 mm/yr (1.47 ft per century). There are many indications that the rate of rise is increasing and will increase during the next century (Titus and Narayanan, 1995; Carton *et al.*, 2005; among others). Miller *et al.* (2001), from work in the marsh behind Hog Island, stated that salt marshes "will be the first ecosystem to feel the effects of an increased rate of sea-level rise." Indeed the interplay between the rates of vertical accretion of the sediment surface and of local sea level is a critical factor in determining the viability of the marsh. When the rate of sea-level rise exceeds the ability of the marsh surface to keep pace, the marsh begins to fail. Oertel *et al.* (1989) found that sediment was accumulating at 36 to 57 percent of the rate of local sea-level rise in marshes on the lower Virginia Eastern Shore. Erwin *et al.* (2006) stated that near Wachapreague and Mockhorn Island "marsh surface elevations do not seem to be keeping pace with local sea-level rise." The postulate that submergence and flooding of the marshes could have major negative consequences on the value of the area as habitat. It is interesting to note that these two studies depict marsh loss, while Finkelstein and Ferland, working in the same area but two decades earlier, reported marsh growth instead of marsh loss. The salt marshes of the coastal zone of Virginia are, at best, on the cusp between survival and failure. Studies designed to monitor the situation would be beneficial as they could provide both a warning and the impetus to plan for the changes that would accompany massive loss of the marshes.

The increased amount of open water could affect the amount of wave energy that reaches the mainland shore of the back-barrier lagoons. The loss of marsh area likely would decrease the mitigation of storm surge, as experienced in southern Louisiana with Hurricanes Katrina and Rita. The increased volume of intertidal open space accompanying marsh loss would result in an increased tidal prism flowing through the inlets, which would yield changes in the cross sectional area of the inlets, and affect sediment dynamics of the inner shelf. Future studies could, monitor, assess, and predict these changes.

The Virginia chain of barrier islands is essentially pure, untouched in any geologically significant way by man. As such, the islands are a natural laboratory for determining how barrier islands respond to physical forcings. Two of the major consequences of global warming are sea-level rise and an increase in the frequency and severity of tropical storms. These are two of the four phenomena that determine the impact of storms on the coast Hobbs (1970). Barrier islands can respond in any of several ways to these mechanisms. They can remain in place and drown as the sea rises around them; they can roll shoreward in some sort of quasi-equilibrium as storm-driven overwash processes carry sand across the island and sea-level rise translates the lagoonal shoreline farther inland; or they can be overtopped by storms and essentially disappear if there is not enough time between destructive events to allow recovery of the shoreface. The third option appears to be situation in the northern group of barrier islands on the Virginia Eastern Shore as the naturally sediment-starved Wallops, Assawoman, Metompkin, and Cedar Islands diminish in elevation and above-sea-level width. It appears that the rate of erosion along Parramore Island, immediately south of Cedar, is increasing. Is this in response to a progressive, downdrift exhaustion of mobile sediment in the nearshore or to a change in the rate of sea-level rise?

As previously noted, the barrier islands of the lower Delmarva Peninsula can be separated into three morphological groups and the boundaries between the groups are coincident with major changes in the morphology of the underlying Chesapeake Bay Impact Crater. Determining whether there is a causative relationship or if it is coincidence would be beneficial in assessing other potential affects of the crater on the regional geology.

The response of the southern Virginia shoreline to sea-level rise also needs to be considered. Will the barrier beach extending south from Sandbridge drown in place as the Atlantic Ocean and Back Bay rise? What would change if an inlet should open somewhere along the northern portion of the long barrier spit? Is the local rate of sea-level rise too great to be mitigated by beach nourishment?

Surveying the Quaternary Sediments of the Middle and Outer Shelf

Although a fair amount of work has been conducted on the inner shelf, within about 20 km (12 mi) of shore, the Quaternary deposits of the middle to outer shelf off Virginia are largely unknown. Wright (1995) discussed the late Pleistocene and Holocene history of the Mid-Atlantic shelf in terms cumulative duration of exposure to potential bottom-disturbing processes. The areas shallower than 20 m have had relatively short exposure to the higher-energy environment of shoaling and breaking waves and strong longshore currents. In contrast, areas of deeper water have experienced resuspension and transport of sediment for a longer time. According to Wright (1995), during the interval between the previous interglacial sea-level highstand (substage 5e) about 125,000 years ago and the present, the shallow portion of the shelf has experienced about 15,000 total years of resuspension events. The cumulative time of exposure increases seaward across the shelf with depth, which suggests that the preserved record offshore (middle to outer shelf) may have a more complex stratigraphy than the nearshore zone. The geologic record of the intermediate stands of sea level (between the high and low extremes), which actually account for most of the time during the Quaternary, are preserved in the stratigraphy of the shelf. Further, the rest of the story associated with the major paleochannels of the Susquehanna-Potomac River system and James River, and their potential connection with the submarine canyons, lies beneath the shelf.

Environmental Studies Concerning Potential Production of Offshore Energy

The generally pessimistic outlook of Bayer and Milici (1987) not withstanding, recently there has been renewed interest in the petroleum potential of the Virginia continental shelf. Should that interest develop to the level of any bottom-disturbing activities, site-specific studies of the near surface geology and sea-floor processes would be prudent. Similar studies would be required for potential drilling sites and pipeline routes.

Rooney-Char and Ayers (1978) concluded that aside from any actual well locations, the site of the landfall would be the most significant element in determining the offshore routing of any pipeline. Their assessments of possible pipeline landfalls broadly considered sites in three areas: the Eastern Shore south of Assateague Island, the bay mouth, and the southern shore. Although they had general conclusions about various sites, the study should be updated as nearly three decades have passed since the work was performed.

The popular press has reported on the early stages of proposals to construct wind farms for the generation of electricity in the offshore regions from Virginia to Maine. One suggestion was for a network of as many as 170 large windmills within a 65 km^2 (25 mi^2) area within 6 km (4 mi) of Cape Charles. The geological and coastal process concerns with constructing such a facility would parallel those for developing the infrastructure for offshore production of petroleum. It would be necessary to consider the characteristics of the subsurface sediments for their ability to support the windmill structure. Sediment movement could affect the stability of structures as could currents and storm waves. As with petroleum pipelines, studies would be necessary for siting the landfalls and offshore routes of the major electrical supply cables. These studies would be in addition to the more obvious studies concerning potential impacts of fisheries, birds, navigation, etc.

Studies Concerning Potential Economic Deposits of Heavy Mineral Sands

Work in the 1980s by Grosz and Eskowitz (1983), Berquist and Hobbs (1986, 1988a, b, 1989), and Berquist (1990) considered the economic potential of deposits of heavy-mineral sands on the continental shelf. These studies basically were "academic" in nature and there was little interest and no participation by industry. However, there are occasional hints of renewed commercial interest. Should that possible interest grow, a two or three stage sequence of studies would be necessary. The first stage would shallow (roughly 5 to 10 m beneath the sea floor) coring to confirm and better define the presence of economic mineral sands and shallow, high resolution seismic profiling to characterize the sand-bodies in which minerals of interest occur. The earlier studies did not have sufficient scope to allow development of a model for the occurrence of the concentrations of heavy minerals. If these studies proved the resource and if there were active, commercial interest in the resource, a suite of studies to assess the range of environmental conditions that might be affected by the dredging and associated activities would need to be conducted. This suite of studies would be similar to the set of environmental assessments relative to sand mining for beach nourishment that have been conducted for the federal Minerals Management Service (Hobbs, 1998, 2000, 2006; among others). The final stage of studies would be related to monitoring the extraction of the resource. Hennigar and Siapno (1990) stated that further (government or government funded) studies should concentrate on gathering data to better define the heavy mineral concentrations so that assessments of the economic/commercial potential of the deposits might be possible.

Updating Sand Inventories

Since the mid-1980s, numerous studies have addressed offshore sand resources on the southeastern Virginia shelf that potentially might serve the ongoing beach nourishment projects at the resort area of Virginia Beach, the U.S. Navy facility at Dam Neck, and Sandbridge (Williams, 1987a,b; Kimball and Dame, 1989; Dame, 1990; Kimball *et al.*, 1991; Hardaway *et al.*, 1995; Hobbs, 1998, 2006; among others). The material used at Sandbridge and Dam Neck has been extracted from Sandbridge Shoal about 5 km offshore. The shoal has been estimated to contain about 20 x 10^6 m^3, from which about 3.5 x 10^6 m^3 had been removed through 2003 (Diaz *et al.*, 2006). Even though this is only a bit over 15 percent of the shoal's volume, there has been no assessment of how much of the shoal could be removed without causing an unacceptable level of environmental change. As it has been two decades since these volumetric assessments were made, and as it is likely that there will be a continuing need to access the offshore sand resources, it soon may be appropriate to update and expand the resource inventories. In addition to revisiting Sandbridge Shoal, the shore-oblique ridges immediately off False Cape should be assessed even though they may be too far offshore to be economically viable. The filled channels that are offshore are another potential resource. If they are carefully mapped in sub-bottom profiles and analysis of core samples indicates that the channel fill is suitable sand, they could prove to be useful reservoirs of sand for beach nourishment or construction aggregate.

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