11. Dynamic Systems at the Land–Sea Interface

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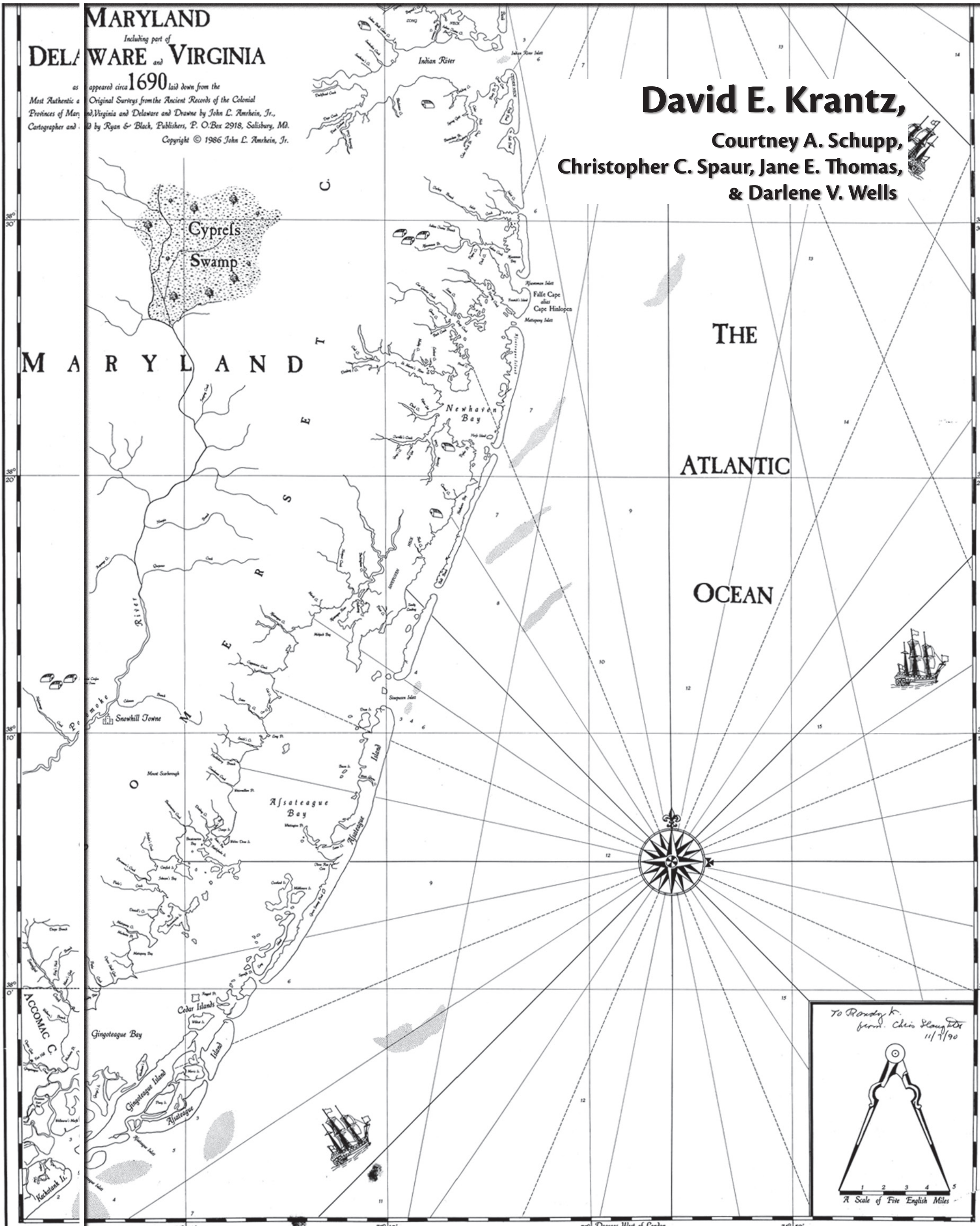


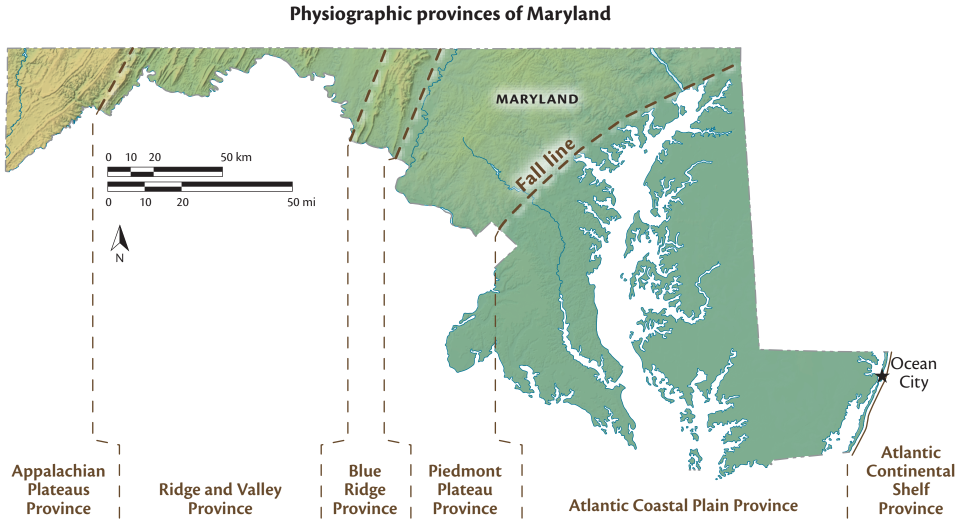
THE PHYSICAL SETTING OF THE COASTAL BAYS

The Coastal Bays are on the Atlantic Coastal Plain

The Maryland Coastal Bays lie east of the mainland of Worcester County, Maryland, adjacent to the Atlantic Ocean and behind the barrier islands of Fenwick Island (Ocean City) and Assateague Island. Stretching from the Maryland–Delaware state line south into Virginia, the bays include Assawoman, Isle of Wight, Sinepuxent, Newport, and Chincoteague Bays. The bays and their Maryland watershed encompass 453 km^2 (175 mi^2) in a narrow strip east of the Pocomoke River watershed. This coastal zone is critical habitat for migratory

Location of Maryland's Coastal Bays

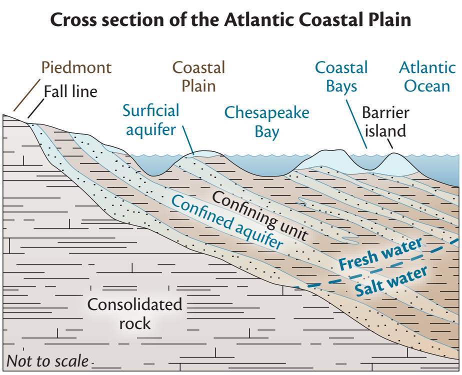
The Maryland Coastal Bays are located on the Mid-Atlantic coast of the United States.



Six regional physiographic provinces run through Maryland, each with a distinct landscape character and geologic history. The Fall Line coincides with a break in slope of the rivers, and marks the boundary between the Piedmont Plateau and Atlantic Coastal Plain provinces. The Coastal Bays lie along the eastern edge of the Coastal Plain.

shorebirds and waterfowl, many important commercial and recreational finfish and shellfish species, and numerous rare and threatened plant and animal species. For more information on the Coastal Bays' living resources, see Chapter 13—*Diversity of Life in the Coastal Bays.*

The Coastal Bays and their watersheds are located along the Atlantic coast of the Delmarva Peninsula on the eastern edge of the Atlantic Coastal Plain province.



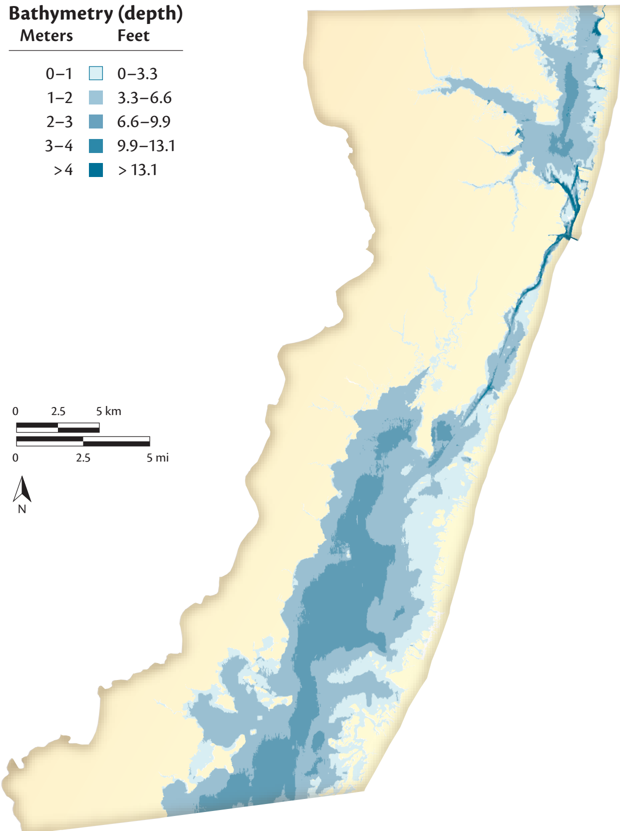
Conceptual cross section of the Atlantic Coastal Plain aquifer system in southern Maryland.[40] The cross section runs approximately northwest to southeast.

The Coastal Plain is characterized by a low and flat landscape underlain by unconsolidated or only partially consolidated sediments such as gravel, sand, silt, and clay. (*Unconsolidated* sediments are those near the land surface that have not been cemented together or compacted by deep burial.) If viewed in cross section, the layers of sediment that underlie the Coastal Plain would appear to be wedge shaped, thinnest in the west and expanding to almost 2.4 km (1.5 mi) thick at the Atlantic coastline.

The Coastal Bays are shallow & flushed by two inlets

The Coastal Bays are shallow, averaging 1–1.2 m (3.3–3.9 ft) deep, except in the navigation channels and inlets. A sandy subtidal flat less than 1 m (3.3 ft) deep created by storm overwash extends as much as 2 km (1.2 mi) westward into the bays from the barrier islands. Wind blowing across the shallow open waters of the bays results in mixing of the water column, meaning that dissolved oxygen levels usually remain high in open-water areas. Surface water input from streams to the Coastal Bays is relatively low because of small contributing watersheds and common sandy soils that allow rapid infiltration of rainwater. Consequently, groundwater is an important source of freshwater inflow to the bays.

Tidal exchange with the Atlantic Ocean is limited to two inlets—the Ocean City Inlet in Maryland and Chincoteague Inlet in Virginia, south of Chincoteague Island. Flushing rates, or water residence times, have been estimated from as short as nine days for Isle of Wight Bay, to as long as 63 days for Chincoteague Bay.[23,37]



Maryland Department of Natural Resources/Maryland Geological Survey.

The Coastal Bays are generally shallow and well mixed.

PROCESSES SHAPING THE COASTAL BAYS

The Coastal Bays are products of modern processes & geologic history

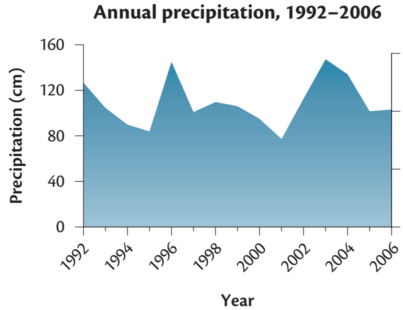
The present Coastal Bays have evolved over approximately the last 5,000 years behind the barrier islands in response to slow, gradual sea level rise. However, in a dynamic coastal environment, the physical factors that sculpt the protecting barrier islands and control the hydrologic characteristics of the bays vary over time scales from twice-daily high and low tides and the two-week cycle of spring and neap tides, up to decadal and centennial changes in the frequency and intensity of coastal storms and wave climate. The greatest changes to the coastal system occur during a few intense storm events that last only a few days but may alter the shape and function of the islands and bays for the next several decades.

Coastal storms have a large impact

Meteorological events such as coastal storms exert considerable influence on circulation, sediments, and water chemistry within the Coastal Bays. The most significant weather effects are brought about by transient cyclonic (low-pressure) or anti-cyclonic (high-pressure) systems. Anti-cyclonic systems, characterized by high barometric pressure and clockwise, outward circulation, usually produce fair, clear weather in Maryland. Cyclonic systems, or low barometric-pressure cells, are the storms. These are more defined, with greater energy and intensity than the anti-cyclonic systems.

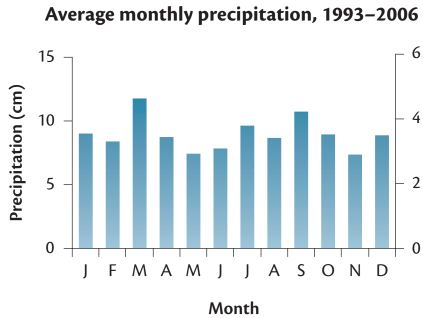
Weather fronts generally move through the region from the west; however, coastal storms have the greatest impact on the Coastal Bays, and the strongest winds from these storms usually blow from the east and northeast. A typical path for extratropical storms—those that form outside of the tropics—is to develop as a low off Cape Hatteras, North Carolina, and then intensify and track northward parallel to the Mid-Atlantic coast. Tropical storms such as hurricanes most commonly approach the region from the south or southwest.

In the summer, mild to moderate prevailing winds driven by the Bermuda High usually blow from the south and southeast, producing gentle waves along the ocean coast and in the Coastal Bays. The most intense summer cold fronts sweeping through the area produce lines of thunderstorms with strong circulation generally from the southwest. Strong southwesterly winds blowing up the axis of Chincoteague Bay can create substantial waves (> 1.2 m or 4 ft, in many cases) that thoroughly mix the water column. In the fall, passing cold fronts often will be



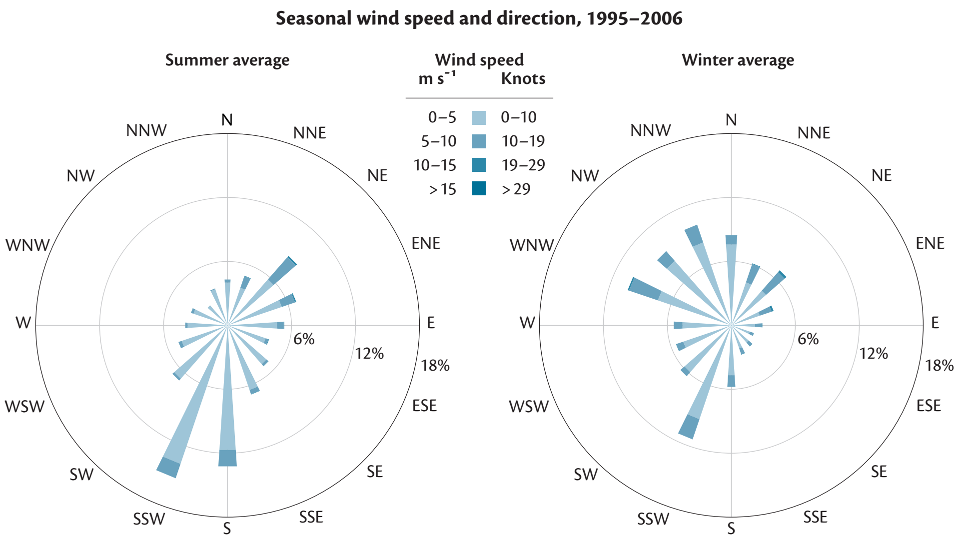
Bureau of Land Management/Assateague Island National Seashore, Remote Automated Weather Station ASTM2.

Precipitation in the Coastal Bays watershed varies year by year. These data are from the weather station approximately halfway along Assateague Island.



National Weather Service.

Monthly mean precipitation for the Maryland coast for the period 1993–2006.



Bureau of Land Management/Assateague Island National Seashore, Remote Automated Weather Station ASTM2.

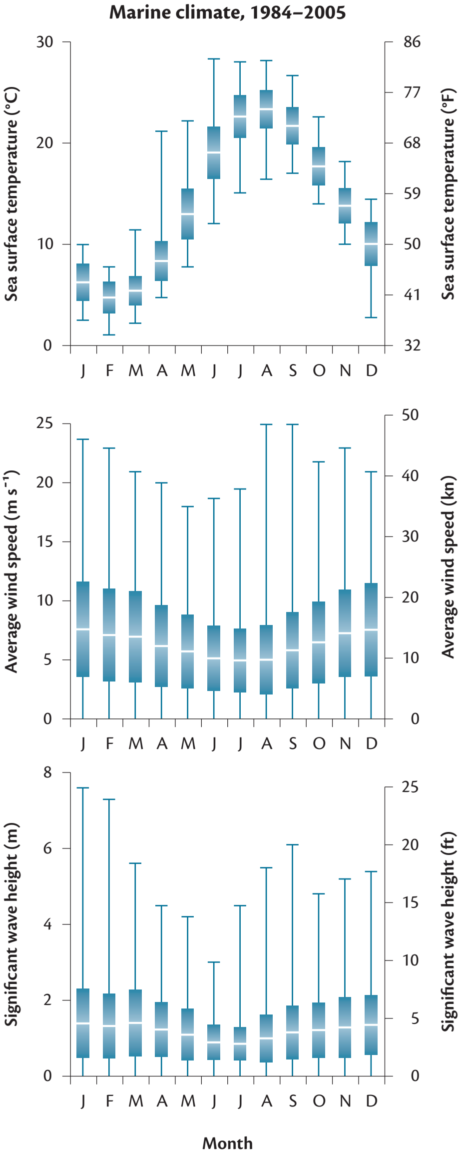
Summer winds are generally out of the southern sector, while in winter, the prevailing winds are from the west. Data are from Assateague Island. For summer, the average was taken from June 21–September 21. For winter, the average was taken from December 21–March 21.

followed by several days of stiff northerly to northwesterly winds generated by the trailing high-pressure system.

Over the past century, the most powerful storms to impact the Maryland coast have been the nor'easter storms. These extratropical storms occur most frequently in late fall (October–November) and early spring (February–March), and generally approach from the south. As the storm moves north along the coast with winds circulating counterclockwise, it produces the highest winds and most damaging waves from the northeast—hence their name, 'nor'easters.'

The wind and wave climate along the Maryland coast has a distinct seasonality. The highest average wind speeds, corresponding to the largest significant wave heights, typically occur in the fall, winter, and early spring. (The *significant wave height* is the average height of the largest one-third of the waves in a particular sea state. These waves contain the majority of the wave energy.) Winds and waves abate, and shift to a more southerly direction, from mid-May through mid-August.

Summaries of two long-term records from instrumented stations off the Maryland coast show that winds of gale force or higher impact the region about one-half of one percent of the time, which is equivalent to two to four storms per year. Offshore waves during these events may rise to greater than 4 m (13 ft) up to



Marine climate for the Atlantic Ocean off the Maryland coast. Data show the median value, first and third quartiles, and lowest and highest values.[32]

Wind speed (knots)	Beaufort scale name	Percentage of observations offshore[1]
0	Calm	1.0
1–3	Light air	5.3
4–6	Light breeze	14.0
7–10	Gentle breeze	25.2
11–15	Moderate breeze	25.4
16–20	Fresh breeze	17.9
21–24	Strong breeze	6.8
25–33	Near gale	4.2
34–47	Gale to strong gale	0.3
> 48	Storm	< 0.1
> 64	Hurricane	≪ 0.1

Annual distribution of wind speed, Atlantic coast of Maryland.

1. NOAA data buoy 44009, 30 km (18.5 mi) east of Fenwick Island at the DE–MD state line in 28 m (92 ft) of water; period of record 1984–2005.

Range of wave height (m)	Range of wave height (ft)	Percentage of observations offshore[1]	Percentage of observations nearshore[2]	Recurrence interval (years) for nearshore[3]
0.0–0.4	0.0–1.3	5.2	31	
0.5–1.4	1.5–4.6	69.5	61	1
1.5–2.4	4.9–7.9	20.0	7.1	2
2.5–3.4	8.2–11.2	3.8	0.9	3–7
3.5–4.4	11.4–14.4	1.1	< 0.1	10–25
4.5–5.4	14.7–17.7	0.2	≪ 0.1	50 or greater
5.5–6.4	17.9–21.0	0.1	—	
6.5–7.4	21.2–24.3	0.1	—	

Annual distribution of wave height, Atlantic coast of Maryland.

1. NOAA data buoy 44009, 30 km (18.5 mi) east of Fenwick Island at the DE–MD state line in 28 m (92 ft) of water, period of record 1984–2005.
2. U.S. Army Corps of Engineers wave gauge MD002, approximately 1 km (0.6 mi) off Ocean City, Maryland, in 9 m (30 ft) of water; period of record 1993–2001.
3. Calculated by U.S. Army Corps of Engineers, refer to *http://sandbar.wes.army.mil*, station MD002.

nearly 7.5 m (25 ft).[32] However, as waves approach shore and move into shallower water, friction with the bottom releases energy and reduces wave height. The beach at Ocean City is hit with waves above 3.5 m (11.5 ft) height about once every 10 years, and extreme waves greater than 4.5 m (15 ft) once every 50 years.[47]

The two storms with the greatest impact on the Maryland coast during the 20th century were a hurricane in 1933 and a nor'easter in 1962. The slow-moving Ash Wednesday nor'easter, as it is known, pounded the entire Mid-Atlantic coast from March 6–8, 1962. This storm coincided with an extra-high spring tide, produced a storm surge of 2.7 m (9 ft) above mean low water, and lasted through five successive high tides. Sustained gale-force winds of 72 km (45 mi) per hour with gusts of 105 km (65 mi) per hour were reported by the Coast Guard Station in Ocean City. This single storm extensively damaged coastal communities from Long Island, New York, through Virginia, and completely reshaped the barrier islands of the Delmarva coast.[45]

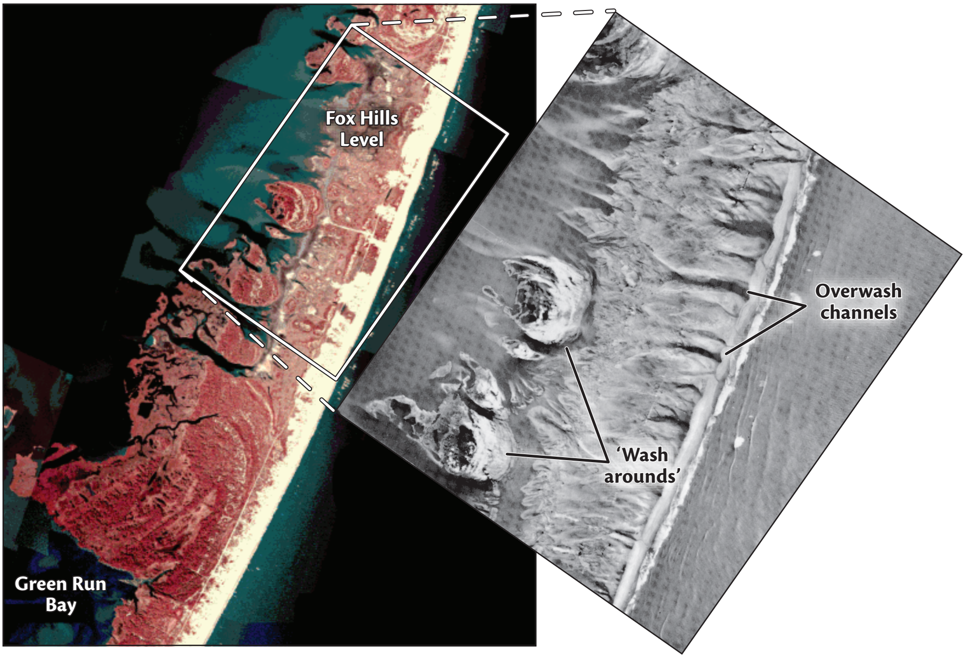
Almost every property in Ocean City suffered some type of damage, from accumulation of sand and debris to total destruction of buildings. Flooding occurred from both the ocean and bay sides, and the elevated water level allowed waves to wash over large sections of

Inlet formed after 1962 storm

National Ocean Service, NOAA.

April 28, 1962. Visible here is the inlet breaching northern Assateague Island that was created by the Ash Wednesday nor'easter storm of March 1962.

Overwash channels and 'wash arounds' created during the 1962 storm



Aerial photo of Assateague Island, showing the overwash channels and 'wash arounds' created by the 1962 nor'easter storm. Photo mosaic at left from Assateague Island National Seashore; aerial photo on right modified.[12]

the barrier islands. Both Fenwick and Assateague Islands were breached in numerous places. Since 1962, Ocean City has not experienced any storms that have approached the intensity and duration of the Ash Wednesday storm.

Unlike extratropical storms, which form in the mid-latitudes during the winter months, hurricanes are cyclonic depressions that originate in the tropical latitudes during the months of July through November. Before reaching the Maryland coast, hurricanes usually lose much of their power and are downgraded to tropical storms or tropical depressions.

Although hurricanes are more severe than nor'easters, they occur less frequently. Studies have shown that during the mid-to-late 20th century, hurricanes crossed the coast (i.e., the storm eye made landfall) between Barnegat, New Jersey, and the southern tip of Assateague Island less frequently than any other area along the U.S. Atlantic coast.[12]

Even though the Maryland coast has suffered few direct hits from hurricanes, even near-misses create large waves and storm surge that drive seawater through the inlets and over the barrier islands into the Coastal Bays.

The hurricane of 1933—which made landfall across the Virginia coast just south of the Chesapeake Bay mouth and moved through Maryland west of Chesapeake Bay—caused the greatest damage to the Maryland coast recorded up to that time. Although of shorter duration than the 1962 nor'easter, the 1933 hurricane was much more intense. Winds during the 1933 hurricane were reported to be 160 km (100 mi) per hour with nearshore waves 6 m (20 ft) high. During the storm, Fenwick Island was breached, forming the current Ocean City Inlet, which was stabilized with jetties soon after the storm and maintained since then by the U.S. Army Corps of Engineers. The stable inlet at Ocean City fundamentally altered the

Coastal Bays ecosystem, especially for Isle of Wight and Assawoman Bays and St. Martin River, which previously were fresher but poorly flushed.

Meteorological events cause storm surge

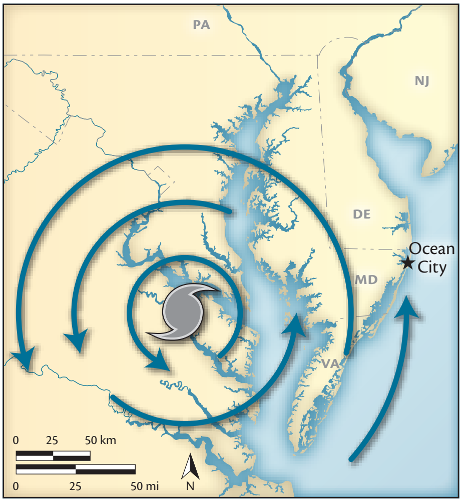
Storm surge—also called a storm tide or hurricane tide—is the rise in water level in the ocean and bays above normal level due to the piling up of water against the coast by the strong winds accompanying a hurricane or intense storm. Reduced atmospheric pressure, particularly the extreme low associated with the center of a hurricane, contributes to the surge. The magnitude of the surge depends on several factors: the size, intensity, and the track and speed of the storm; the shape of the coastline; nearshore topography; and the astronomical tide stage. A storm surge is potentially catastrophic, especially if landfall coincides with high tide and is accompanied by extremely high wind-driven waves. Worldwide, most hurricane deaths are caused by the storm surge.

Storm surge height typically is measured as the sea level height observed during the storm above the normal sea level height. Storm surges of 3–4 m (10–13 ft) are common for moderate, Category 1 or 2 hurricanes. Recent extreme storm surges include a 6-m (20-ft) surge during Hurricane Hugo striking South Carolina in 1989, and the devastating 8-m (26-ft) surge generated by Hurricane Katrina along sections of the Louisiana and Mississippi coast in 2005.

The strongest winds of a hurricane are typically in the northwest quadrant, and a storm moving from southeast to northwest can push a large wall of water (the storm surge) ahead of it. The most damaging hurricane that could impact the Maryland coast would approach slowly from the southeast, creating the maximum surge and striking the coast almost at a right angle. Most Atlantic hurricanes over the last decade stayed out to sea, passing the Maryland coast, or made landfall in North Carolina or Virginia and tracked west of the Coastal Bays.

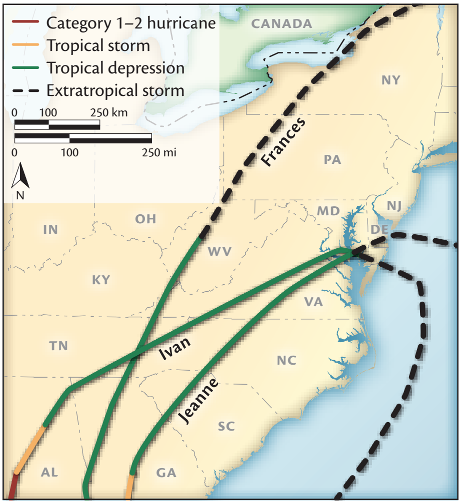
In September 2004, tropical storm remnants of three hurricanes (Frances, Ivan, and Jeanne) all tracked west of the Coastal Bays, but still caused widespread flooding from storm surge. Due to

Storms that track west of the Coastal Bays

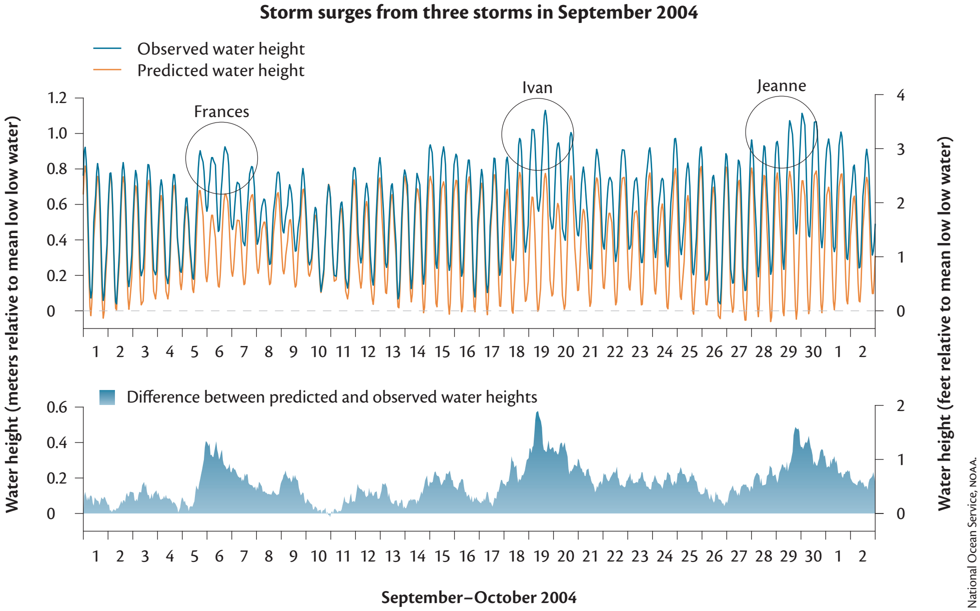


Storms that track west of the Coastal Bays can produce a storm surge by pushing water into the Coastal Bays, due to the counter-clockwise rotation of the storm.

Storm tracks of three 2004 hurricanes



Storm tracks of the three storms—Frances, Ivan, and Jeanne—that tracked west of the Coastal Bays in September 2004.



Storm surge from three storms that affected the Coastal Bays during September 2004. Data are from NOAA tide gauge station #857-0283, located in Isle of Wight Bay at the Ocean City Inlet.

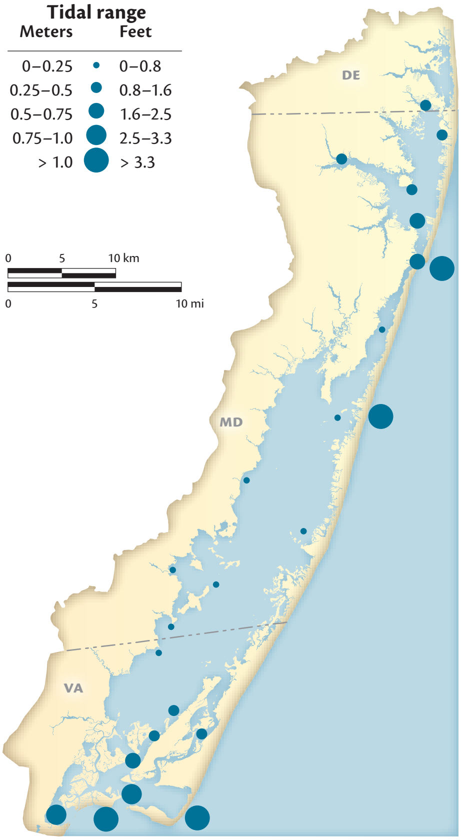
the counter-clockwise circulation of hurricanes, even when they travel west of the bays, strong southerly winds along the Atlantic coast push seawater through the inlets into the bays.

Wind events influence the Coastal Bays more than astronomical tides

The Maryland coast is classified as microtidal, with a tidal range in the ocean less than 1.5 m (4.9 ft). (*Tidal range* is the elevation difference between high tide and low tide.) The tidal range near the two inlets is a little more than 1 m (3.4 ft), and diminishes rapidly with distance from the inlets. The range drops to 45 cm (17.7 in) in Assawoman Bay, and 12 cm (4.7 in) in central Chincoteague Bay. Because the Coastal Bays are shallow with a very small tidal range, tidal currents are relatively weak except near the inlets, and wind events influence the bays more than the astronomical tides.

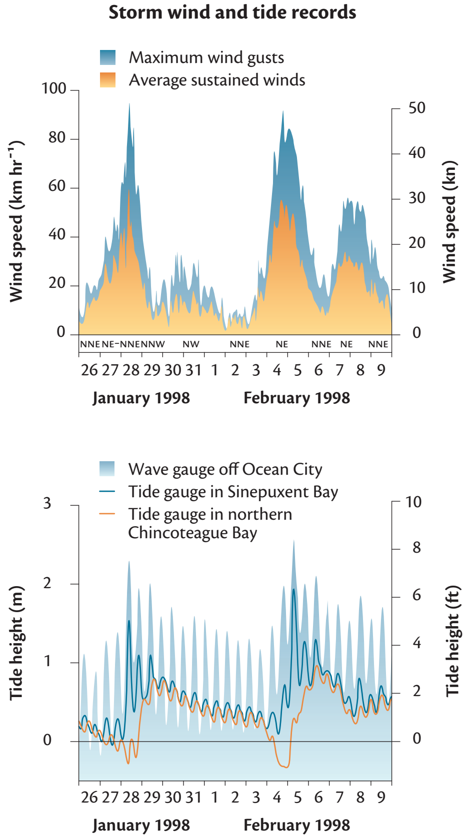
When strong winds blow persistently for a day or more over the open water of the bays, the wind stress, or friction of the wind with the water surface, both creates waves and transports the water downwind. Besides the direction, duration, and intensity of the wind, the *fetch*—or the distance over open water that the wind blows—controls the response of water level in the bays. For example, the long axis of the connected Chincoteague and Sinepuxent Bays trending southwest to northeast is nearly 50 km (31 mi) long, but Chincoteague Bay is only 6–8 km (3.6–4.2 mi) wide in most areas. Consequently, wind blowing from the northeast gains a lot of traction and will push the water in the bay to the south—similarly, southwesterly winds pile up water in northern Chincoteague Bay.

The effect of storm winds on bay water is seen in the wind and water-level records from two nor'easters in 1998. During the first day of each storm, the water level in the ocean and measured at Ocean City Inlet rose above the predicted tidal height by as much as 1.5 m (4.9 ft). However, at the same time, the water level in northern



NOAA, National Park Service, and Maryland Geological Survey.

Tidal range in the Coastal Bays and the adjacent Atlantic Ocean. The Coastal Bays are microtidal.



Remote Automated Weather Station.

NOAA, National Park Service, and u.s. Army Corps of Engineers.

Wind and water levels in the Coastal Bays during two nor'easter storms in the winter of 1998. Modified.[39]

Chincoteague Bay dropped dramatically, as the bay water was transported to the south, and did not rise until the storm passed and the northeasterly winds weakened. Although there are no tide-gauge data from the southern part of Chincoteague Bay for these events, the water levels most likely rose by at least 0.5 m (1.6 ft) during the peaks of the storms.

These wind events enhance the otherwise sluggish exchange of water between the Coastal Bays and the coastal Atlantic Ocean as a large volume of ocean water is driven through the inlets into the bays during the storm, and bay water exits to the ocean as the storm recedes. Wave action in the bays produces opposing effects on water chemistry—air bubbles in white caps and breaking waves oxygenate the water column, but at the same time anoxic sediments are stirred up from the bay floor, releasing chemically reduced compounds and nutrients such as ammonium and phosphorus.

Winds between 20 and 33 knots—stronger than a fresh breeze but below gale force—occur about 11% of the time along the Maryland coast and may be associated with the passage of weather fronts or low-pressure systems. Typically, these moderately strong winds, especially if sustained for more than six hours from a constant direction, are capable of producing whitecaps and churning up fine sediments from the bay floor over most of the Coastal Bays; however, there have been few systematic studies of these processes in the Coastal Bays. The high turbidity from these wind events may persist for one to two days after the winds abate. When accompanied by elevated water levels, these more-frequent, moderately strong wind events also contribute to the erosion of tidal marshes by wave attack along the outer edge of the marsh and undercutting of the root mat.

Barrier islands protect the Coastal Bays

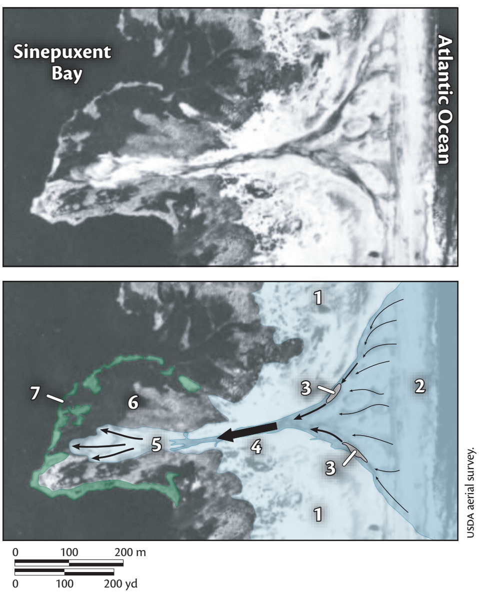
Assateague and Fenwick Islands protect the Coastal Bays and mainland Maryland from major storms and some of the impacts of sea level rise. The barrier islands absorb much of the energy of waves and currents produced by storms before they reach the bays and mainland. However, in extreme storms, the combination of elevated water level and large waves in the ocean result in the overwash of sections of the barrier island. Low-lying areas of the islands are particularly susceptible and may be breached to form a new tidal inlet. A view of central Assateague Island following the 1962 storm shows the after-effects of 1–2 m (3.3–6.6 ft) of seawater driven by waves flowing across Fox Hills Level. Nearly 45 years later, the channels, storm ridges, and 'wash-arounds' created by that event are still visible on the island.

During major storms, large quantities of sand from the beach and shoreface (the innermost shelf) are transported over the island and deposited on the back-barrier flat and in the bays. Overwash may be extensive, as seen after the 1962 storm, or more localized in overwash fans. Wind blowing across the emergent back-barrier overwash flats before plant cover is re-established may rework the sand to create low dunes on this otherwise remarkably flat surface. Over time, the overwash process builds up the height of the island and creates the shallow platform for tidal marshes and subtidal flats along the eastern edge of the Coastal Bays.

On time scales of hundreds to thousands of years with gradual sea level rise, overwash and the opening and closing of inlets allow the barrier islands to roll over and migrate landward. Development on Fenwick Island and construction of artificial dunes has effectively prevented overwash and sand deposition on the bay side of the island since the mid-1970s.

The Fenwick Island beach has been maintained by the state and U.S. Army Corps of Engineers since the late 1980s through beach nourishment measures. Sand dredged from offshore shoals in the Atlantic Ocean now provides the source of sand for longshore transport along Fenwick Island. About 610,000 m^3 (800,000 yd^3) are placed on Fenwick Island from these sources every four years.[48]

Overwash system on Assateague Island



1. Extent of overwash from most recent storm.
2. Catchment area for channelized overwash.
3. Deeply scoured section of overwash channel.
4. Convergence of overwash flow.
5. Distributary flow and deposition to form fan.
6. Washover fan from previous storms.
7. Reworking of washover sands by waves.

An individual overwash system on northern Assateague Island created by a coastal storm.

GEOLOGY & HYDROLOGY OF THE COASTAL ZONE

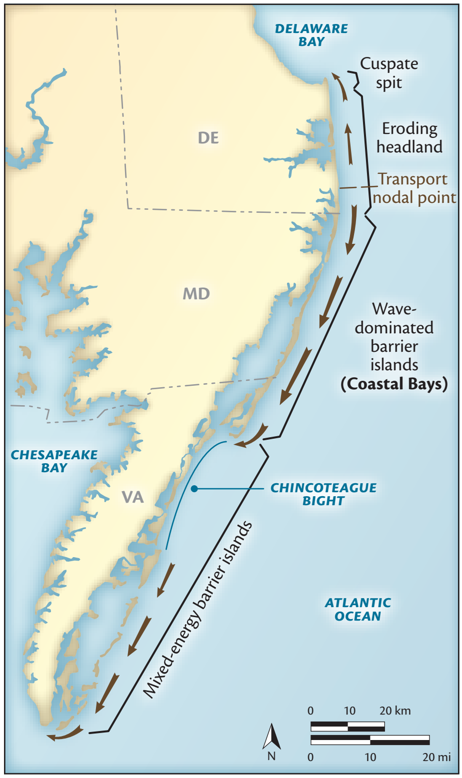
Sand moves along the coast

The regional transport of sand along the coast affects the overall morphology of the barrier islands, and local variations create hot spots of erosion. Wave action moves sand grains along the beach in the swash zone (area of surf run-up) and, more importantly, drives currents flowing parallel to the beach that carry sand grains. The annual net transport of sand is a balance between energetic flow to the south generated by nor'easter storms during the winter, and less vigorous but more constant flow to the north produced by southeasterly waves during the summer.

During average years with moderate storms, sand in the upper shoreface (the inner shelf near the beach) is transported in a natural cycle. Vigorous waves during storms (primarily in winter) move sand offshore to form bars, and fair-weather summer waves push sand back onshore to widen the beach. For the Maryland coast, 6–6.5 m (20–21 ft) seems to be the depth limit for this offshore–onshore transport of sand.[41] However, during major storms, bottom sediments over most of the shelf are set in motion as large, long-wavelength waves churn the water column. During these storms, the zone of longshore sand transport may extend offshore 1–2 km (0.6–1.2 mi) into water as deep as 20 m (66 ft).

The Mid-Atlantic coast may be subdivided into several *coastal compartments* that are bounded by the mouths of major estuaries and have similar geomorphic character:[15] Long Island, New Jersey, the Delmarva Peninsula, and the Virginia–North Carolina coast from Cape Henry to Cape Lookout. For the entire Mid-Atlantic coast, the net longshore transport of sand is to the south or southwest. In the northern section of each coastal compartment, the net transport reverses in response to gradients

Delmarva coastal compartments



The Atlantic coast of Delmarva (Delaware–Maryland–Virginia) may be subdivided into four coastal compartments, each with a characteristic morphology related to sediment transport (brown arrows) and wave and tide energy.[15,34] The Chincoteague Bight is an embayed section of the ocean coast south of Chincoteague Island and the southern tip of Assateague Island, where the barrier islands are offset landward as much as 10 km (6.2 mi).

in wave energy and shoreline orientation downdrift of the estuary mouth.

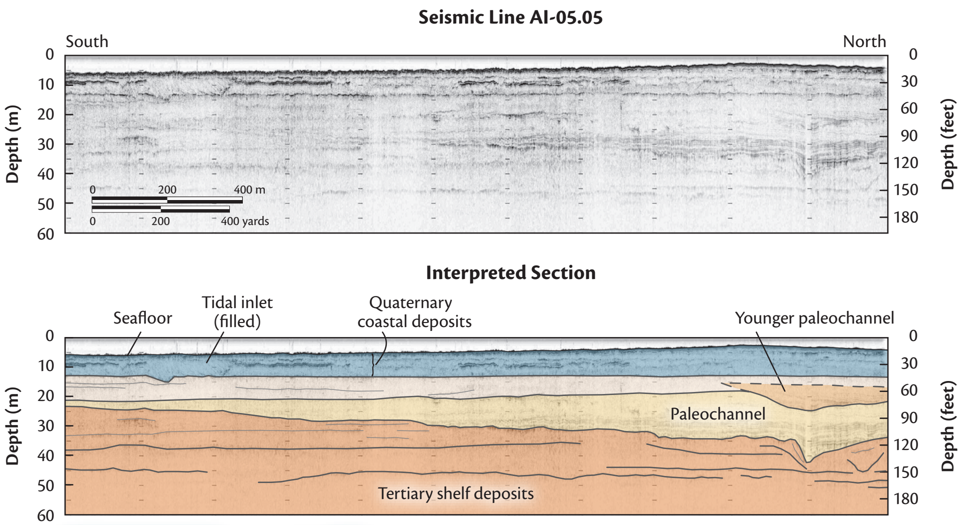
The dividing point between zones of northerly and southerly transport is known as a *nodal point*. Within each zone, local reversals of transport direction occur near inlets and with major changes in shoreline orientation. The nodal point along the Delmarva coast lies between South Bethany, Delaware, and the Delaware–Maryland state line.[8]

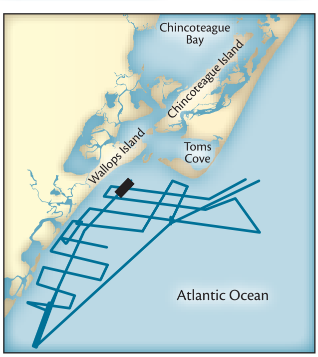
For the Delmarva coast south of the Delaware–Maryland state line, net longshore transport is to the south and is estimated to be between 115,000 and 214,000 m^3 (150,000–280,000 yd^3) per year and increases to the south.[46] Local reversals south of the Ocean City and Chincoteague Inlets transport sand northward.

Engineered structures such as groins and jetties intercept longshore transport, accumulating sand on the updrift side but starving the downdrift side and enhancing erosion. Nodal points form at the sites where transport diverges. For example, at South Bethany Beach, Delaware, transport is both to the north into the mouth of Delaware Bay and to the south toward Fenwick Island.[8] This divergence is in part due to the sheltering effect of Cape May Shoals and Hen and Chickens Shoals at the mouth of Delaware Bay. Storm waves approaching from the northeast lose energy by breaking across the shoals and refract around the shoals, changing the direction at which they hit the Delaware coast north of Bethany Beach. This same effect happens on a smaller scale downdrift of inlets and results in a curved or arc-shaped offset of the shoreline.

The southern end of Assateague Island in Virginia terminates in a large hook, or recurved spit, that bends sharply to the west and wraps around Chincoteague Island. This feature, comprising Toms Cove Hook and Fishing Point, receives sand transported southward from the entire coastal compartment that stretches south from the nodal point at South Bethany, Delaware.

The Virginia barriers of Wallops, Assawoman, and Metompkin Islands are offset landward by almost 10 km (6.2 mi)





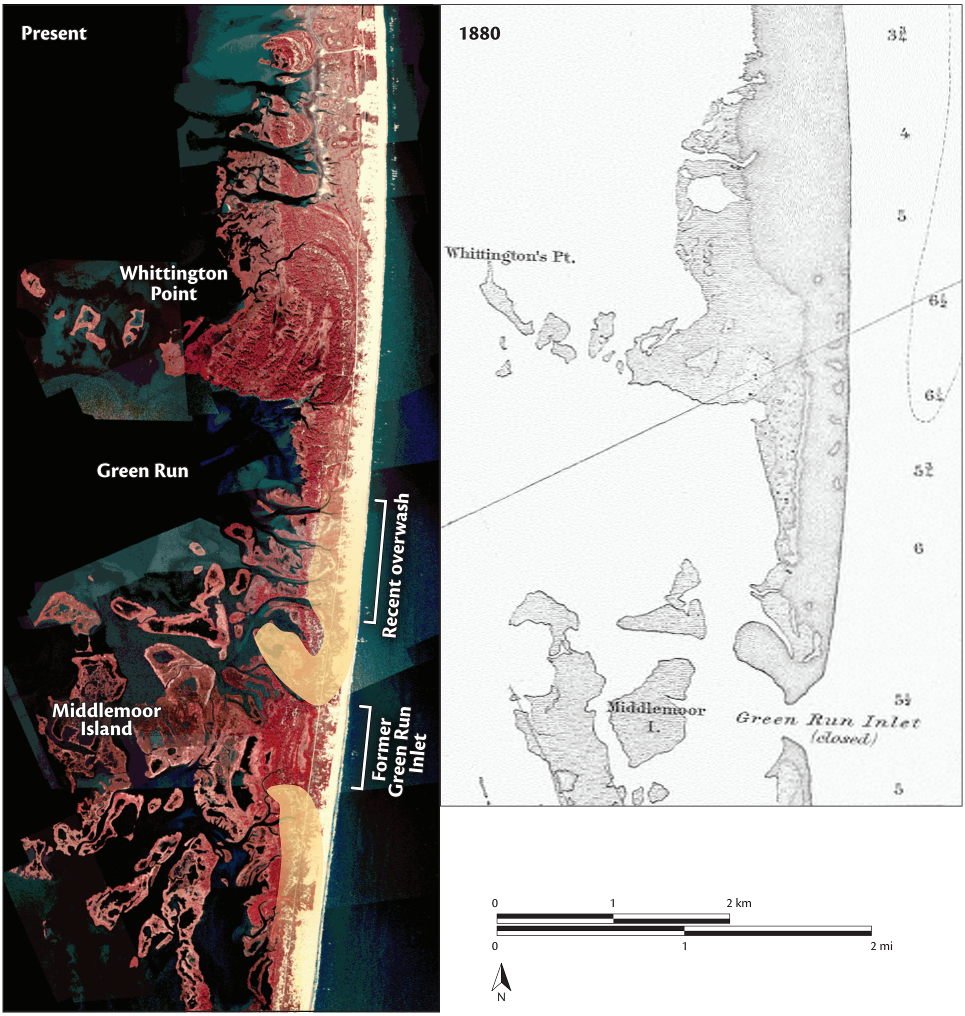
Several paleochannels cut into Tertiary shelf deposits were discovered in a recent seismic survey of the Chincoteague Bight.[21] These may be channels of an ancestral Potomac River that flowed across the shelf to Washington Canyon prior to the creation of Chesapeake Bay.

from Assateague Island, forming an embayed area referred to informally as the Chincoteague Bight. The recent discovery during seismic surveys of the Bight of several large, deep paleochannels, possibly of an ancestral Potomac River, explains the presence of what appears to be a broad valley extending across the shelf toward Washington Canyon. This feature appears to have persisted through many sea level cycles and may be the underlying reason for the offset and change of character of the barrier islands south of Assateague Island.[21]

Islands roll over & inlets migrate

Viewed over several centuries, these same processes of storm overwash and creating and filling inlets result in island

Location of the former Green Run Inlet



The aerial photo (left) shows the location of a former inlet through Assateague Island at Green Run in Chincoteague Bay, and the nautical chart from 1880 (right) shows the inlet mapped as a shoreline feature. The former flood tidal delta—Middlemoor Island—is visible. Photo mosaic on left from Assateague Island National Seashore; historical chart from NOAA National Ocean Service.

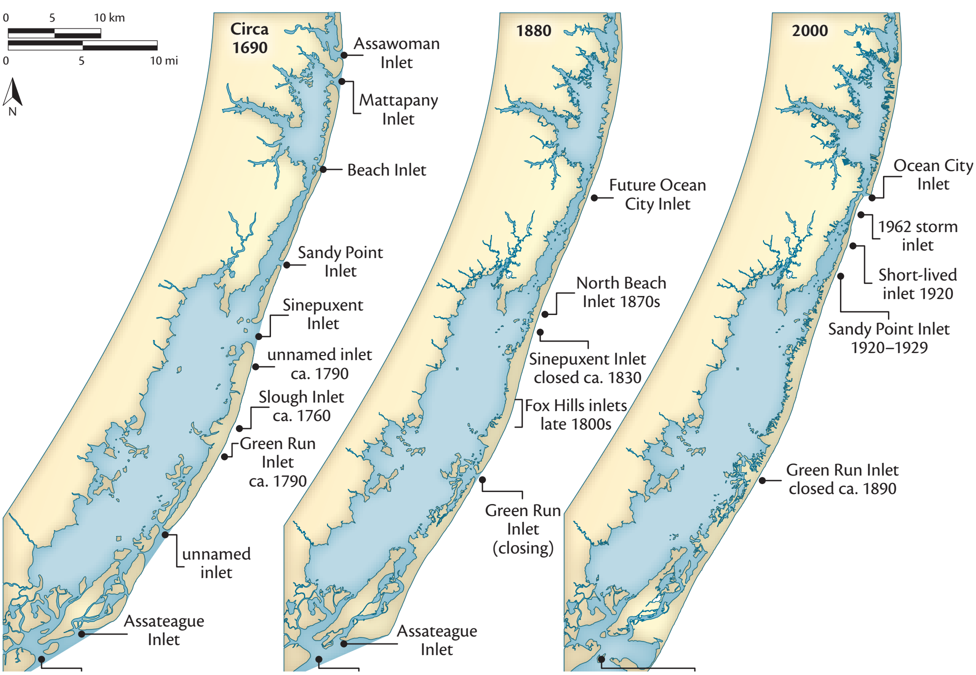
rollover and landward migration (coastal transgression). An inlet cut during a storm may remain open for years or decades if the *tidal prism*—the volume of water in the bay between high and low tide—is large enough to generate vigorous tidal flow that scours sediment and maintains the inlet.

However, the Delaware–Maryland coast is microtidal (tidal range of less than 1.5 m or 4.9 ft) and wave-dominated. Consequently, sand transported along the ocean shoreline from north to south usually fills and closes off the new inlet within a few years. Historical maps and charts going back to the first European colonists in the late 1600s show the sequence of inlets opening and closing along Fenwick and Assateague islands.[1,27]

Natural coastal processes prevail along the length of Assateague Island beyond the influence of the jetties at the north end. Even so, since the closure of Green Run Inlet in the 1890s, only Sandy Point Inlet (1920–29) remained open for any substantial length of time. Further, since construction of the jetties at Ocean City Inlet in 1933, no new inlets have formed other than at the north end. The reasons for this relative stability of Assateague Island are not entirely clear, but include the buffering of storm surges by flow through the two inlets, the construction in the 1960s of an artificial foredune along much of the length of the island,[24] and subsequent vegetative cover by grasses and scrub behind that dune.

Even under natural conditions, inlet creation and closure will affect the stability of that section of the island for several decades. This effect can be exacerbated by human activities. For example, the northern end of Assateague Island is classified by the U.S. Geological Survey as having 'very high vulnerability' to sea level rise because of its low elevation, frequent overwashing, and high rates of shoreline erosion.[36]

Inlets on Fenwick and Assateague Islands, 1690–2000



The shoreline of Fenwick and Assateague Islands in 1690, 1880, and 2000, showing the locations of various inlets.

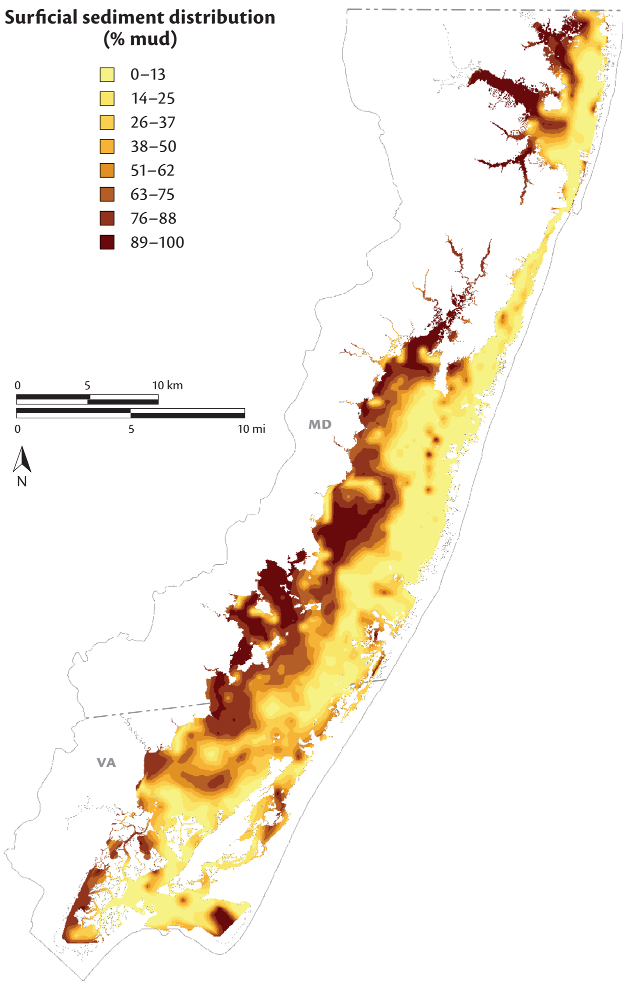
The underlying cause for this instability is the interruption of the longshore transport system by the jettied Ocean City Inlet, and enhanced capture of sand by the ebb shoal and other accretionary shoals well beyond the capacity of what natural features would have had at this location. Downdrift sediment starvation results on northern Assateague Island.

Wave & tide energy determine sediment type in the Coastal Bays

The two dominant sediment types in the Coastal Bays are medium to fine sand (particles with diameters of 0.062–0.5 mm) and organic-rich mud (a watery mixture of silt and clay with particles less than 0.062 mm diameter). The distribution of sediments in the bays is related to the proximity of the sediment source and the wave and tide energy available to transport the sediment particles. The tidal inlets, with deep channels and complex flood-tidal deltas, are composed almost entirely of sand with some gravel, although they grade into silty fine sand on the bayward end. The overwash sheets on the back-barrier flat also are dominantly sand, but commonly will be interbedded with marsh peats and lagoonal silts deposited between storm events.

Away from the inlets and channels, the Coastal Bays essentially are a large settling basin for fine-grained sediments. In Chincoteague Bay, sands dominate the eastern section, silts fill the navigation channel, and silt–clay mud covers the western one-third. The protected upper reaches of the bays, such as Newport Bay, the western tributaries of St. Martin River, Turville Creek, and Greys Creek, have minimal wave and tide action, and clays are deposited there.

Sediments deposited in the bays are derived from several local sources—suspended sediments transported by the streams, turbid seawater churned up on the shelf during storms, erosion of the



Sediments in the Coastal Bays range from sandy sediments along the eastern margins, to finer sediments along the landward shores.

mainland shoreline, and overwash across the barrier islands. A minor contribution comes from windblown sand lifted off the islands.

Sediments are habitat for the benthic community

The sediments of the bay floor of the Coastal Bays provide the substrate for the benthic community of plants and animals. Vastly different benthic communities occupy the different areas of predominant sediment type. This is due largely to the different geochemical environments, including concentrations of nutrients and dissolved oxygen, associated with sandy versus finer-grained sediments.

The silt and silt–clay sediments in the central and western sections of the Coastal Bays typically are dark gray to black, organic-rich, and disaerobic to anaerobic (little to no dissolved oxygen) within a few millimeters of the sediment–water interface.[52] The sediment pore waters have high concentrations of nutrients, usually as ammonium (NH_4^+) and phosphate (PO_4^{3-}), released by the remineralization (decomposition) of organic matter, and methane gas produced by bacterial methanogenesis.[4] Because of the abundant organic matter, these sediments support a highly productive benthic invertebrate fauna. These sediments also are an important component of seasonal cycling within the Coastal Bays, acting at different times as both source and sink for nutrients and contaminants. For more information on nutrients, see Chapter 12—*Water Quality Responses to Nutrients*.

The resurgence of seagrasses (submerged aquatic vegetation or SAV) in the Coastal Bays since the 1980s[49] has been attributed to recovery from a wasting disease that devastated the grasses in the 1930s and improving water quality since the 1970s (although this resurgence trend has recently reversed). Although the habitat criteria monitored are associated with water quality (nutrient

Jane Thomas.

These overwash flats on the landward side of Assateague Island supply sand to Sinepuxent Bay, creating habitat for seagrasses.

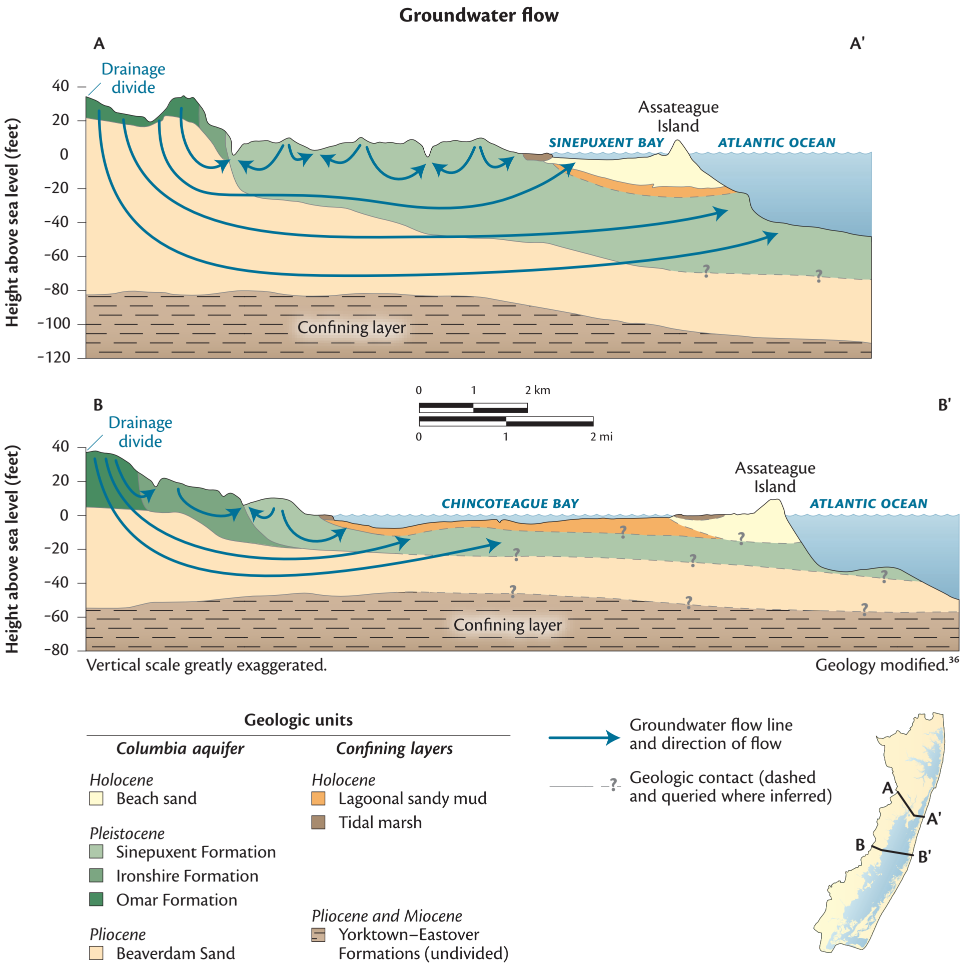
concentrations, Secchi depth, suspended solids), a sandy substrate appears to be a necessary condition for healthy seagrass beds. Nearly 85% of the seagrasses grow on the sandy subtidal overwash flats and flood-tidal deltas on the landward side of Assateague Island, including the two large deltas associated with previous inlets at Green Run and north Chincoteague. Most of the remaining seagrasses live on local patches of sand near shorelines where waves are eroding sandy Pleistocene deposits, such as at Mills Island and South Point in Chincoteague Bay. For more information on seagrasses, see Chapter 14—*Habitats of the Coastal Bays and Watershed*.

Stream drainage into the Coastal Bays is limited

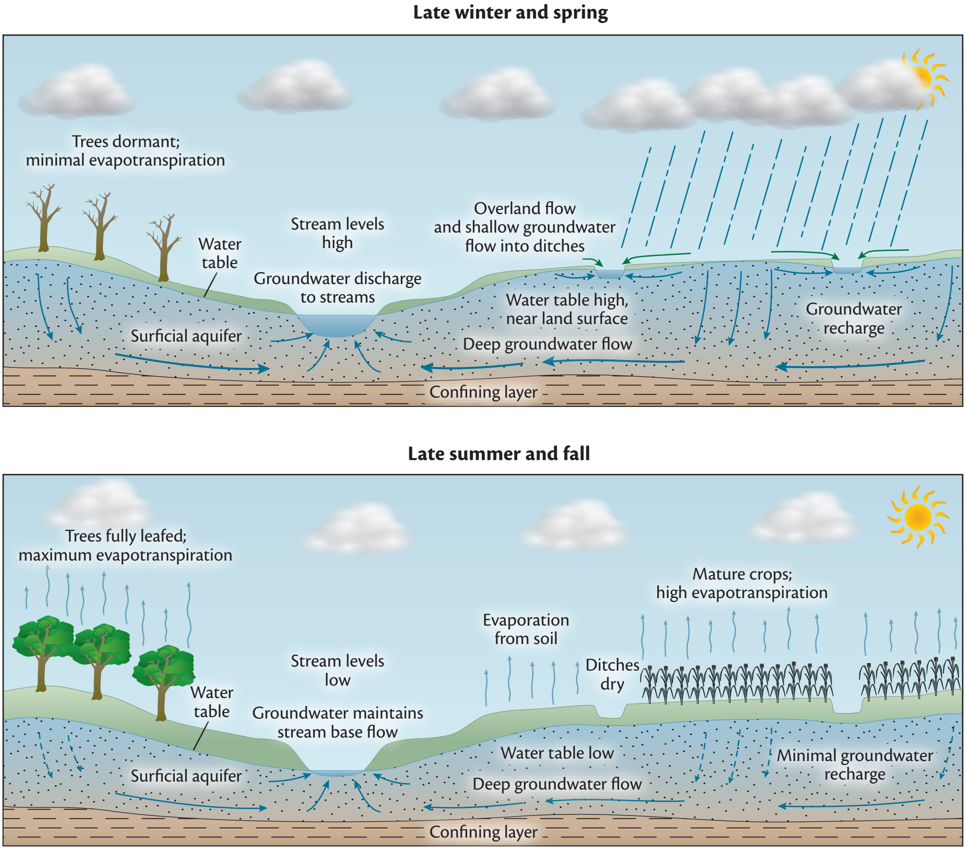
The morphologic character of the Coastal Bays varies along the length of the Delmarva coast, with substantially more drainage area contributing to the Delaware coastal bays than to either the Maryland or Virginia bays. Rehoboth and Indian River Bays in Delaware flood two fairly substantial incised valleys (the stream drainage network created during the last glacial lowstand of sea level). The watershed extends well inland and drains much of eastern Sussex County, Delaware. In contrast, the watershed contributing to Chincoteague Bay is made up of numerous small streams that extend the short distance to the drainage divide with the Pocomoke River and have a relatively low discharge of freshwater. The drainage landward of the lagoons behind the Virginia barrier islands is even more restricted, and these bays have the lowest ratio of drainage basin area to estuary area along the Delmarva coast.

Groundwater flows into & under the Coastal Bays

Because much of Delmarva has sandy sediments near the land surface, as much as 25–50% of annual precipitation infiltrates through the soil and recharges the water table.[2] This groundwater, which fills the pore spaces between sediment grains, then flows through the surficial, or water-table, aquifer. An *aquifer* is simply a body of sediment (or rock) that has sufficient permeability to allow water to flow through easily—these typically are continuous layers of sand in the subsurface. The surficial aquifer is open to the land surface, but deeper aquifers may be isolated by low-permeability layers (confining layers) composed of silt or clay that inhibit vertical water flow. Many municipal water supplies, including those for Ocean City and Salisbury, Maryland,



Local and regional flow of groundwater in the surficial aquifer of the Coastal Bays watershed. Modified.[11]



Groundwater flow and recharge changes depending on the season and precipitation.

draw from these deeper confined aquifers.

Groundwater in the surficial aquifer may follow a short flow path and discharge into a nearby stream or may sink deeper and enter a longer, regional flow path. During periods with no rain, and no overland runoff, stream flow is maintained by groundwater discharging upward through the stream bed—this is *stream base flow*. Analogous to the watershed comprising the stream drainage network, there is an equivalent ground-watershed contributing to base flow.

On Delmarva, much of the *groundwater recharge*—addition of new water from precipitation to the water table and surficial aquifer—occurs during the winter and spring, when evapotranspiration is low, precipitation is high, and the water table rises to near the land surface.[2] At this time, in ditched areas such as farmland, numerous short, shallow flow paths drain off groundwater into ditches and streams. Conversely, in summer and early fall, high rates of evapotranspiration remove moisture from the soil and unsaturated zone, the water table drops dramatically, and ditches and shallow streams dry up.

Groundwater flow in the vicinity of the Coastal Bays or other estuaries is complicated by the salinity, and thereby higher density, of the surface water. Similar to the mixing of riverine freshwater with seawater in an estuary, the dynamics of freshwater and saltwater mixing are mimicked in the subsurface beneath the bays. In general, areas close to the mainland shore will have fresh groundwater flowing out toward the bay, with the denser saline groundwater that recharges through the bay floor sinking and flowing landward beneath the fresh. In many coastal settings, this interaction produces a zone of fresh groundwater discharge that extends tens to a few hundred meters (up to 300–500 ft) into the bay, with an underlying wedge of saline groundwater pushing landward.

The balance between fresh and saline groundwater at the shore of the Coastal Bays depends on the hydraulic head generated by recharge on the mainland—think of the pressure produced by the height of a water tower that pushes water through the supply pipes and out a faucet—but also depends` on the stratigraphy (type and layering of sediments). Modeling predicted that groundwater recharged on the mainland would flow beneath a narrow Coastal Bay, such as Sinepuxent Bay, and discharge offshore in the ocean.[11] Conversely, in wider bays such as Chincoteague Bay, which is 10 km (6.2 mi) wide in parts, the fresh groundwater meets significant resistance to flow from the saline groundwater recharged from the bay, and is either forced upward buoyantly to discharge through the bay floor or, more likely, mixes to form a brackish 'subterranean estuary.'[30]

Field observations in Indian River and Rehoboth Bays in Delaware, and Chincoteague, Sinepuxent, and Isle of Wight Bays in Maryland, using a marine electrical resistivity system towed behind a small boat, showed a complex interaction between fresh and saline groundwater beneath the bays[20,25] and generally verified model predictions.[11] Plumes of fresh or nearly fresh groundwater extend 1–2 km (0.6–1.2 mi) from shore under sections of the bays beneath thin semi-confining layers. The groundwater within and adjacent to these plumes was sampled by drilling temporary wells through the bay floor to evaluate the chemical characteristics and age of the water.[4,5]

The Coastal Bays that separate the barrier islands from the mainland also isolate the shallow groundwater of the island from fresh groundwater flowing from the mainland. Long, narrow barrier islands such as Assateague and Fenwick Islands will have a relatively shallow lens of fresh groundwater literally floating on saltwater flowing under the island from both the ocean and the bay. From field observations, the fresh lens on Assateague Island is 6–7 m (20–23 ft) thick in the middle of the island, and pinches to less than 1 m thick toward the ocean beach and the bayside marshes. The fresh lens is isolated vertically from the deeper fresh groundwater flowing from the mainland by the layer of saltwater.[16]

SEA LEVEL & THE EVOLUTION OF THE COASTAL BAYS

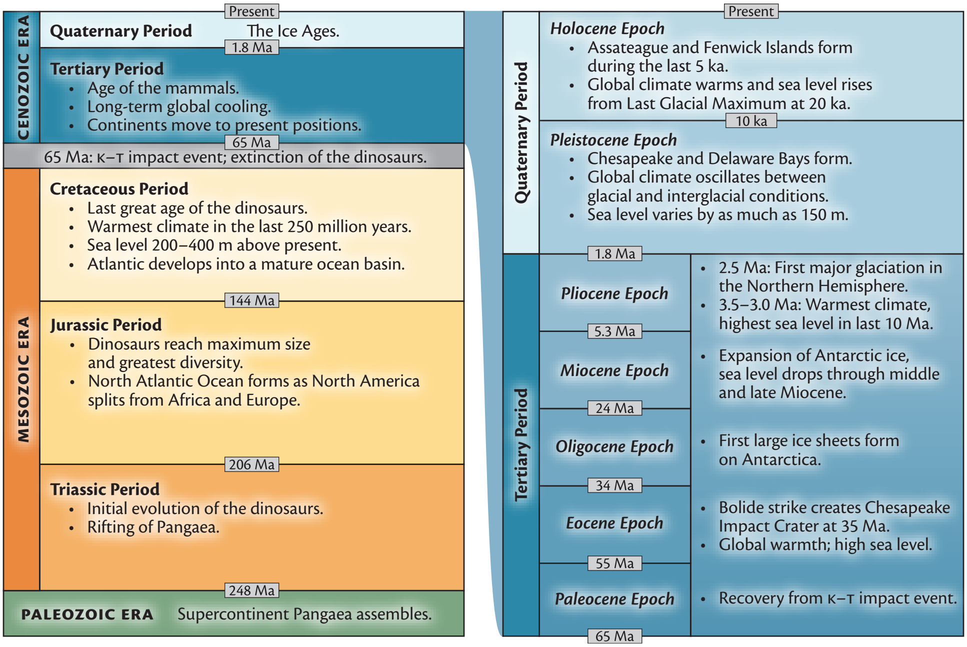
The Coastal Bays inherit a geologic legacy

While the modern processes shape the character of the Coastal Bays, the bays are a product of their geologic history. This legacy developed as the Atlantic Coast of North America evolved over 250 million years (Ma, or 'mega annum') and sea level has varied by nearly 150 m (500 ft) in glacial–interglacial cycles during the past 2.5 Ma.

The sediment layers of the Atlantic Coastal Plain record a long and complex history from the initial rifting of the supercontinent Pangaea in the early Triassic to form the Atlantic Ocean basin, through flooding of the margin by the ocean and deposition of marine shelf sediments on the edge of the continent, to the most recent rise of sea level since the last major Ice Age 20,000 years ago (ka, or 'kilo annum').

The Delmarva Peninsula as we now know it began forming during the Pliocene and early Pleistocene (from about 5–1.5 Ma) as the ancestral Potomac

Geologic timeline



The major subdivisions of geologic time for the Mesozoic and Cenozoic Eras are shown along with important events affecting global climate and sea level over the past 250 million years. ("Ma" is million years, "ka" is thousand years.)

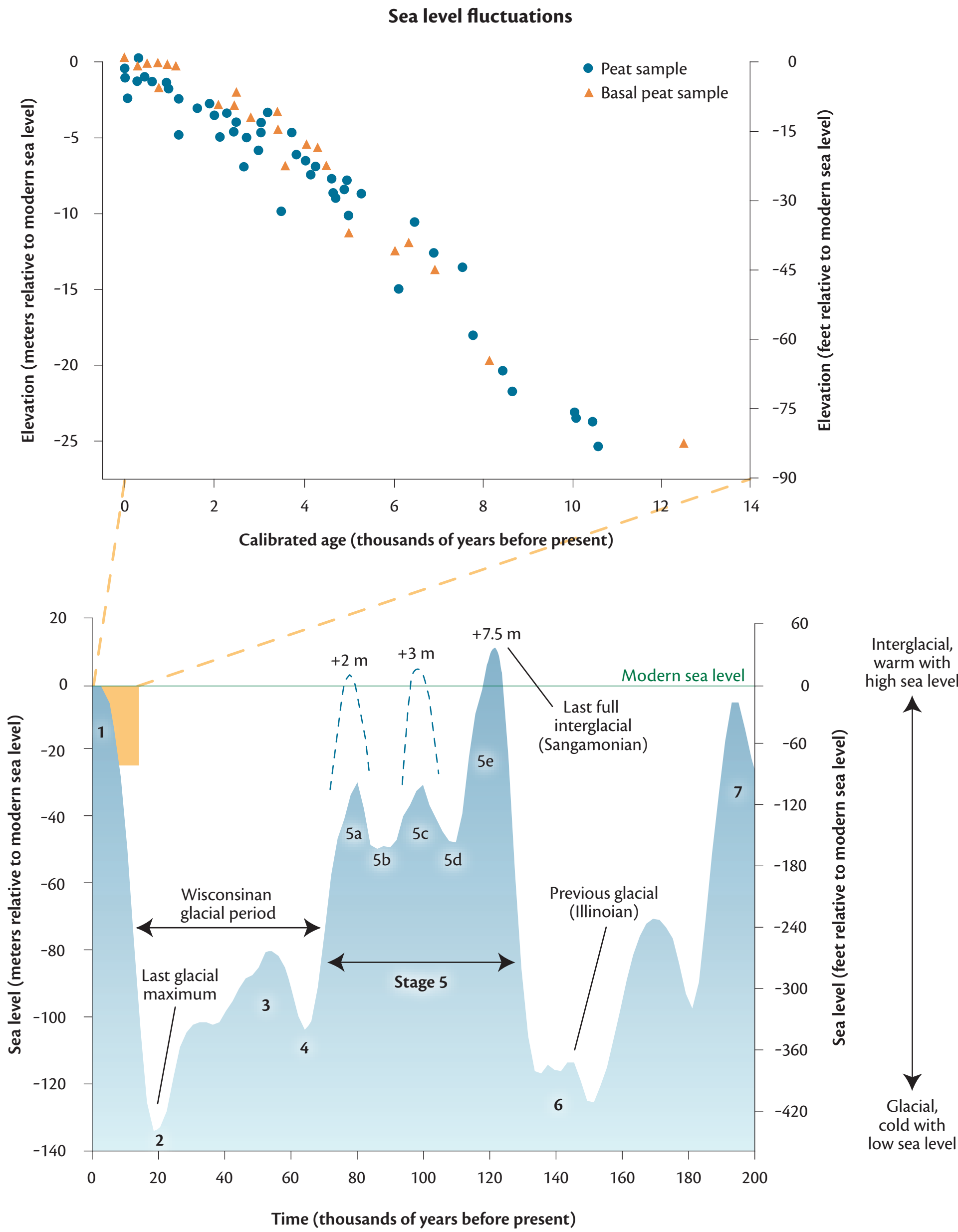
and Delaware Rivers deposited deltas and braided-river outwash plains that would become the core of the peninsula. Although continental glaciers did not extend this far south, many of the sediments deposited on Delmarva were derived from glaciers scraping off the land surface of Ontario, Quebec, Pennsylvania, and New York, and dumping large volumes of water and sediment into the major river systems.

Since approximately 2.5 Ma, global climate has oscillated between cold glacial and warm interglacial conditions. During glacial episodes, the Antarctic ice cap grows and ice sheets as thick as 3–4 km (1.8–2.4 mi) cover most of Canada and Scandinavia. The water stored in these continental ice sheets is removed from the oceans, and sea level drops by 120–150 m (400–500 ft).

During the glacial lowstands of sea level in the Middle Pleistocene (1–0.5 Ma), the Delaware and Susquehanna Rivers started cutting down through the Coastal Plain to create the basins for Delaware Bay and Chesapeake Bay, respectively. During the interglacial highstands of sea level, the ocean flooded landward across the Coastal Plain (presently the continental shelf), and wave action eroded the deltaic sediments deposited previously and redistributed them into sandy shorelines. As sea level dropped with the beginning of the next cooling phase, the highest shorelines would be stranded, creating long linear features that can be traced most of the length of the Delmarva coast.[10,29,44] The steep seaward face of these shorelines is a *scarp*, and the flat plain between scarps is a *terrace*.[7,29]

Prominent shorelines mark previous highstands of sea level

The most recent interglacial period with sea level as high or higher than modern occurred between 125 and 80 ka—this is referred to as the Sangamonian interglaciation or Marine Isotope Stage 5. As seen in the sea level plot for the past 200 ka, Stage 5 had a stair-step of three highstands at approximately 125 ka (substage 5e), 100 ka (5c), and 80 ka (5a). The peak sea level of substage 5e rose to 6.5–7.5 m (21–24.5 ft) above present, producing a prominent shoreline along most coasts around the world. Records



NOAA National Climate Data Center.

Sea level fluctuations globally and locally along the Mid-Atlantic coast over the past 200,000 years. The lower panel shows global ice volume converted to sea level relative to present.[26] Dashed blue lines indicate local elevations on Delmarva of sea level highstands associated with substages 5a and 5c. The upper panel shows the rise in sea level over the last 14,000 years for the Delaware coast from radiocarbon dates of peats and wood.[38]

of global ice volume for substages 5c and 5a indicate that these two subsequent sea level highstands should have risen only to about 20 m (65 ft) below present. However, shoreline deposits from these events are found between 2–3 m (6.5–10 ft) above present along the Delmarva coast and elsewhere in the Mid-Atlantic. This apparent anomaly may be explained in part by depression of the Mid-Atlantic shelf by the weight of seawater, thus allowing sea level to rise relatively higher in the region.[50]

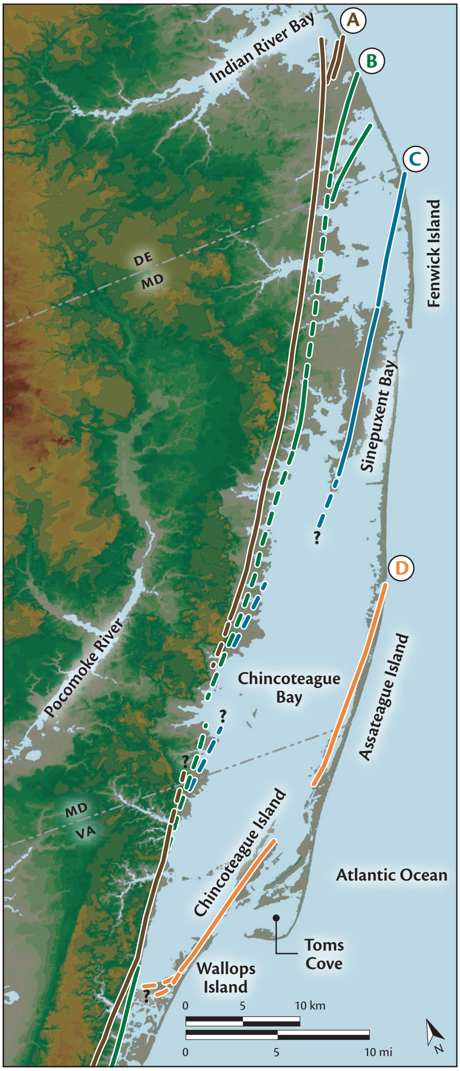
The last three Pleistocene shorelines are most important for the development of the modern Coastal Bays. The highest and most prominent shoreline, created at 125–115 ka during substage 5e (shoreline 'A' on the map), can be traced from Cedar Neck on the south bank of Indian River Bay in Delaware[9] to the Joynes Neck Sand on the Atlantic side of Cape Charles, Virginia,[29] and correlates with the Ironshire Formation in eastern Worcester County, Maryland (shown on the cross section on page ??).[35] The lower-elevation and younger shorelines of substages 5c and 5a (~100 ka and ~80 ka) are associated with subtle lineations of the Bethany (Delaware)/Sinepuxent Neck (Maryland)/Wachapreague–Bradford Neck (Virginia) complex,[44] immediately west of the Coastal Bays (shorelines 'B' and 'C' on the map).

From the preceding warm period, global climate cooled and progressively falling sea level bottomed out at about 125 m (410 ft) below present at 22–20 ka during the Last Glacial Maximum of the Wisconsinan glaciation. At this time, the modern continental shelf was emergent as a broad, exposed coastal plain, and the shoreline was at the edge of the shelf. Locally, the stream drainage network incised to create the basins that would be flooded to form the Coastal Bays.

Sea level rose as glaciers melted

Deglaciation, or the melting of the continental glaciers, proceeded slowly at the start, but accelerated by about 18 ka. The water from the melting glaciers flowed back into the ocean and sea level

Late Quaternary shorelines



The topographic relief produced by late Pleistocene shorelines partially controls the shape and character of the Coastal Bays. The trends of late Quaternary shorelines are shown on a color digital elevation model for the Delmarva coast. Late Pleistocene shorelines produced during highstands of sea level are correlated with periods of reduced global ice volume: [A] substage 5e, ~125–115 ka; [B] substage 5c, ~105–95 ka; and [C] substage 5a, ~85–75 ka. The Wallops–Chincoteague–Pope Island shoreline trend [D] may have been produced by a sea level event in the late Holocene.

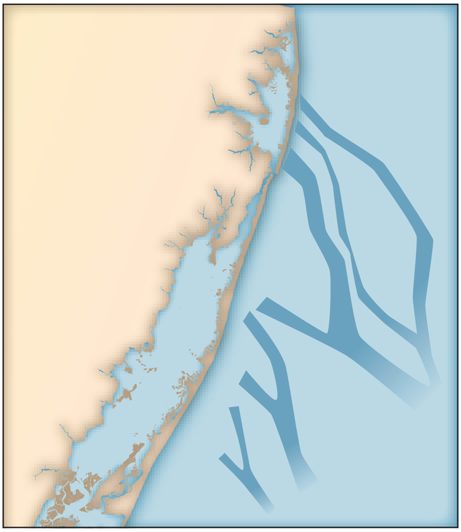
rose. During some intervals between 15 and 10 ka, sea level was rising 10 times faster than today. The barrier islands that existed on the edge of the shelf during the lowstand migrated rapidly landward across the flooded continental margin. Sea level rise slowed by about 5 ka, and the barrier islands moved close to their present positions.

The Coastal Bays formed behind barrier islands

As sea level rose, the low-lying areas between the barrier islands offshore and the Pleistocene shoreline ridges on the mainland flooded to create the Coastal Bays.

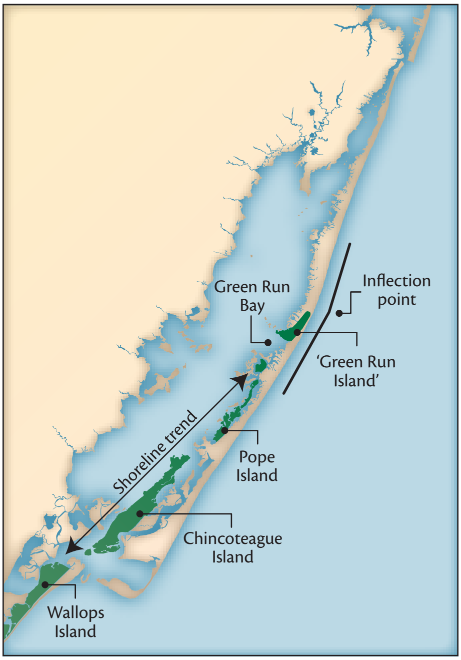
Possibly around 2,500 years ago, although no radiocarbon dates are currently available, a precursor to Assateague Island formed slightly west of the modern island. This shoreline complex

Ancestral stream valleys



Stream drainage networks cut during the last Ice Age lowstand of sea level (including the ancestral St. Martin River) are now filled and covered by shelf sediments. Modified.[43,51]

Late Holocene Chincoteague shoreline



Shoreline complex showing the precursor to Assateague Island to the west of the current position of the island.

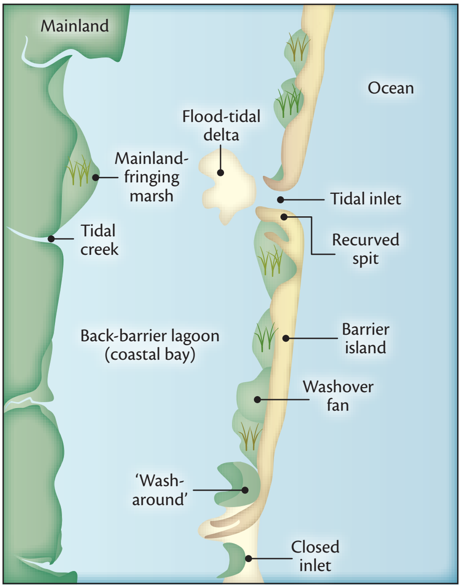
can be seen today as a linear trend that includes the western part of Wallops Island, Chincoteague Island, Pope Island, and Green Run Island—the latter two sections lie behind Assateague Island near the Maryland–Virginia state line.

This older coastline has a morphology similar to the barrier islands of the Virginia Eastern Shore, with short island segments separated by relatively stable tidal inlets—a mixed-energy coast instead of the wave-dominated modern Assateague Island. Expansive flood-tidal deltas and sets of beach ridges on the north end of Chincoteague Island and at Green Run mark the locations of former inlets and suggest that the inlets were stable for long periods of time, possibly several hundred years.

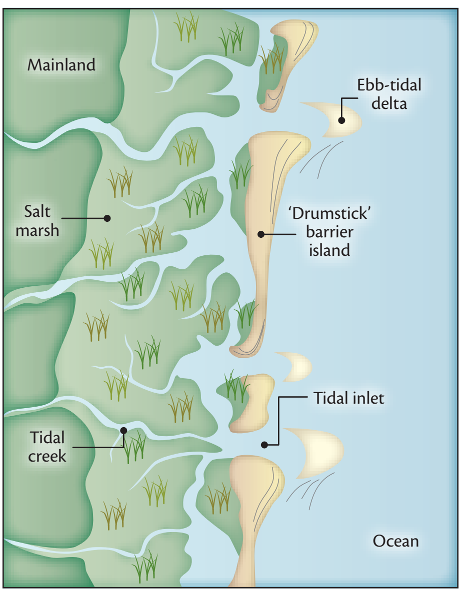
Even in this initial phase of development, the individual bays were distinctly different because of their geologic heritage. The St. Martin River drainage is the largest of the Maryland coastal drainages, and the incised valley is wider, deeper, and extends farther inland than the other stream valleys.

Consequently, Isle of Wight Bay (including the St. Martin River estuary) is more similar to Indian River Bay in Delaware than the other Maryland Coastal Bays.

Microtidal, wave-dominated barrier island



Mixed-energy (wave and tide) barrier islands



The morphology of barrier-island coasts varies depending upon the balance between wave and tide energy.[17] The Coastal Bays fall into the top category, being microtidal and wave dominated.

Assawoman Bay formed as the valleys of smaller tributary creeks to St. Martin River flooded; however, this basin was confined to the east by the extension of the late Pleistocene Sinepuxent Neck shoreline that trends under Fenwick Island near the Delaware–Maryland state line and creates the easternmost point of the Delmarva coast. Newport Bay, Sinepuxent Bay, and two small estuarine tributaries, Herring Creek and Ayers Creek (west of Sinepuxent Neck) formed in a trellis drainage network controlled by the coast-parallel trend of the Pleistocene shorelines, notably Sinepuxent Neck.

Because of the large, stable inlets, the ancestral Chincoteague Bay would have been different in character from the modern bay. The dynamics were probably more similar to the marshy coastal lagoons behind the Virginia barrier islands, with greater influence of seawater and exchange with the coastal ocean, more fine suspended sediment entering the Coastal Bays from the ocean, substantially larger tidal range and more vigorous tidal currents, and considerably shorter water residence time. If a new inlet were to form and stabilize in central Assateague Island, the dynamics of the Coastal Bays most likely would revert to this former character.

Sea level is still rising

The amount of water on the surface of the planet is finite and constantly circulated from the sea, through the atmosphere, to the land, before eventually being returned to the sea—this is the *hydrologic cycle*. The major storage components in the hydrologic cycle are the oceans, groundwater, and continental or glacial ice—lakes and rivers account for a small percentage of the total water volume.

As discussed above, the amount of water in the oceans fluctuates as global ice volume increases or decreases, causing a proportionate fall or rise of sea level, respectively. Over the past 2.5 million years, repeated episodes of glacial expansion and melting have corresponded with quasi-periodic falls and rises of global sea level. For the past 500,000 years, the major glacial lowstands and peak interglacial highstands of sea level have recurred with an approximate 100,000-year periodicity.

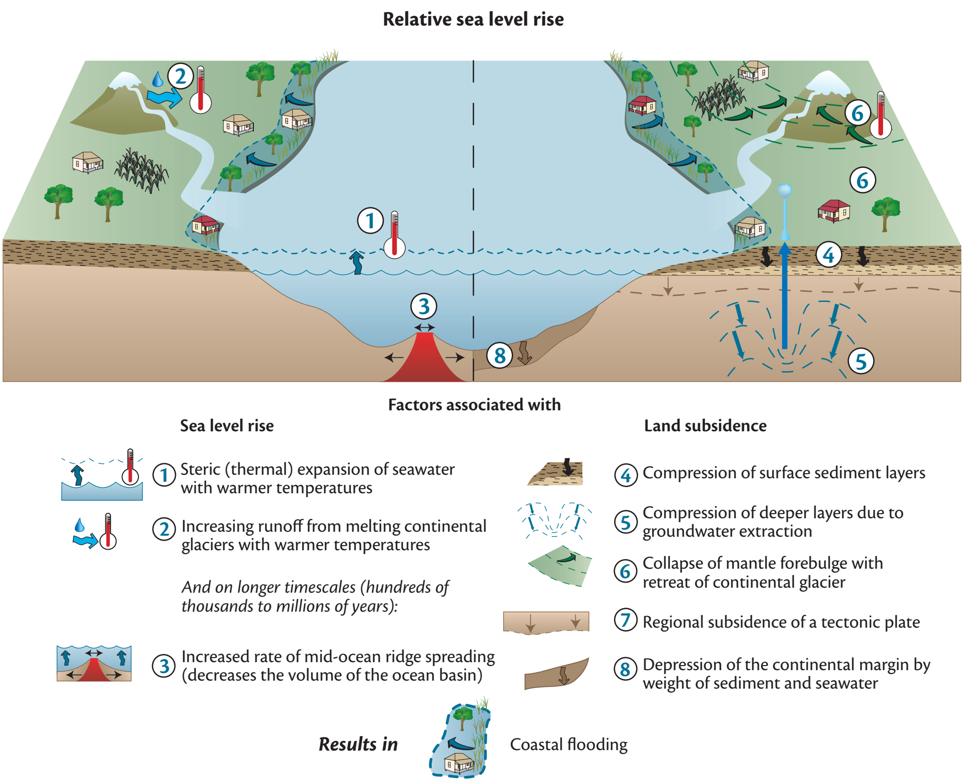
Relative sea level is a combination of sea level rise & land subsidence

Sea level has been rising for the past 18,000 years, although not at a constant rate. Global sea level during the 20th century rose at a rate of 2 mm (0.08 inch) per year, but this rate is nearly ten times that of the previous several millennia.[13,28]

Global warming contributes to sea level rise in two main ways—more runoff from melting ice sheets and glaciers enters the ocean, and as the water in the surface ocean (the upper 500 m or 1,650 ft) warms, the volume of the water increases from thermal expansion.

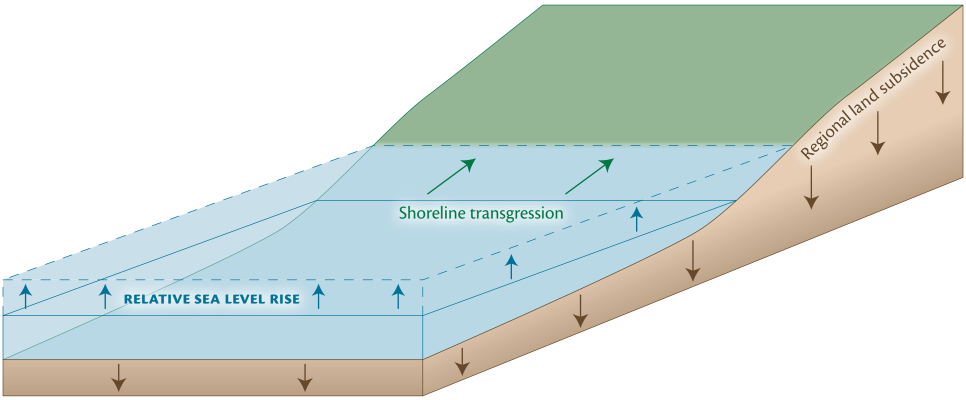
For a coastal area, the relative sea level is a combination of the global sea level rise and local or regional land subsidence. Relative sea level rise along the Delmarva coast during the 20th century ranged from about 3–4 mm (0.12–0.16 in) per year[33,53] or 30–40 cm (12–16 in) per century, but varied spatially with local geologic and hydrologic conditions.

Regionally, the position of the continent relative to sea level will change depending upon the history of loading and unloading of glacial ice onto the continent, or the flooding of the continental margin by the ocean. This process is *isostatic adjustment*, and it occurs because the hot rocks of the upper mantle are plastic and can flow slowly when uneven force is applied. For example, the rocky coast of Maine, which



Relative sea level is a result of a combination of factors.

Land subsidence and sea level rise



In the Mid-Atlantic region, regional land subsidence is a major factor contributing to relative sea level rise.

was depressed several hundred meters by the weight of glacial ice, has been rebounding, or uplifting, since the ice retreated.

During the maximum glaciation, the weight of the ice on North America depressed the crust under the ice, and created a bulge of mantle material around the edge of the ice—similar to pressing down on a closed tube of toothpaste. During the glacial period, the Mid-Atlantic region sat high astride this forebulge. As the ice sheet melted and retreated, the forebulge collapsed and migrated northward, and the land surface in the Mid-Atlantic subsided, even as the Maine coast rebounded.[31]

Throughout the last 15,000 years, in addition to the forebulge collapse, the ocean has been progressively flooding the Mid-Atlantic continental margin. As with glacial ice, the weight of the added seawater depresses the crust. The subsidence from these combined effects contributes the extra 1–2 mm (0.04–0.08 in) per year to the relative sea level rise for the Delmarva coast.

Rising sea level has myriad effects

Rising sea level has both regional and global consequences because of the potential to alter ecosystems and habitability of coastal regions, where an increasing proportion of the world's population lives.

Increasing sea level can result in coastal erosion, exacerbated flooding and storm damage, inundation and loss of wetlands and other low-lying areas, salt intrusion into aquifers and surface waters, and higher water tables.[13] Higher sea-surface temperatures associated with global warming are likely to increase the frequency and intensity of tropical storms such as hurricanes.[3]

The position and shape of barrier islands will change through time in response to sediment supply, changes in predominant wave direction and energy, and rate of sea level rise.[42] A barrier island is stable when the rate of sea level rise is approximately balanced by the input of new sediment to build up the island. In the case of substantial sediment input or relative sea level fall, the front of the island will build seaward, or prograde. This seaward *regression* will produce a series of beach ridges called a strand plain.

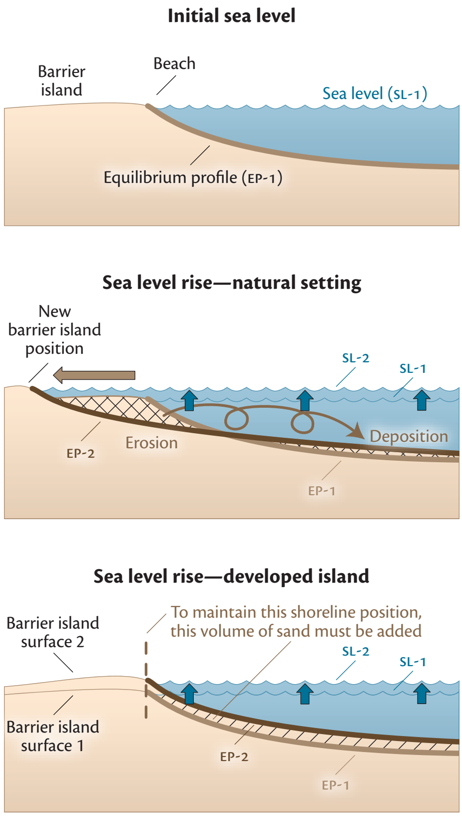
With a slow rise of sea level, the barrier island will retain its general morphology but will roll over and migrate landward, which is called *transgression*. However, in the case of rapid sea level rise, or an interruption of sediment supply, the transport processes that restore sand to the island cannot keep up with the rate of landward migration. These conditions result in low, narrow barrier islands that are vulnerable to storms and are frequently washed over and cut by ephemeral tidal inlets, similar to the north end of Assateague Island, or Metompkin and Assawoman Islands in Virginia.

Sea level rise may accelerate in the next century

A recent report by the United Nations Intergovernmental Panel on Climate Change[19] (IPCC) stated that the global average rate of sea level rise of 2 mm per year (about eight inches per century) that prevailed through most of the 20th century has increased to approximately 3 mm per year (1 ft per century). The same IPCC report also noted that global warming increased faster than expected in the past decade. Current predictions of global sea level rise for the coming century range from 50–90 cm (1.6–3 ft),[18,19] but have a degree of uncertainty because of complex interactions among global climate and the hydrosphere and cryosphere (global ice).

During the 20th century, sea level along the Delmarva coast rose by as much as 40 cm (16 in), or roughly twice the global average. The resulting landward retreat of the ocean shoreline was about 20 m (66 ft),[53] although this was not uniform for the entire coastline. Because the Mid-Atlantic continental margin has such a low slope—generally less than one degree of inclination—for each centimeter of sea level rise, the shoreline migrates landward 50–120 cm (1.6–3.9 ft).[53] Coastal engineers commonly use the 'Bruun rule'[6] to estimate the step-back of the shoreline.

The basic concept of the Bruun model is that a shoreface will have a parabolic 'equilibrium profile' that is a balance between available sediment and dissipation of wave energy. When sea



Bruun[6] developed the concept of an 'equilibrium profile' for the shoreface, which has a parabolic shape caused by the gradual release of wave energy approaching shore. This model can be used by engineers to predict the landward movement of the shoreline with sea level rise and to estimate the volume of sand required for beach nourishment projects.

level rises, the profile is translated upward and landward, resulting in erosion and transport of sediment from the beach. In a natural setting such as Assateague Island, this results in island roll-over as described previously.

When faced with continuing shoreline erosion, some coastal communities with mostly small residential buildings are adopting a 'strategic-retreat' policy in which ocean-front structures are removed from their foundations and moved landward. However, in the case of a barrier island developed with large hotels and condominiums, such as Ocean City, moving multi-story buildings is not feasible. The alternative is to add sand to the beach, in approximately the volume indicated by the difference between the old and new equilibrium profiles. Concurrently, artificial dunes are constructed and stabilized to minimize overwash and flooding during coastal storms.

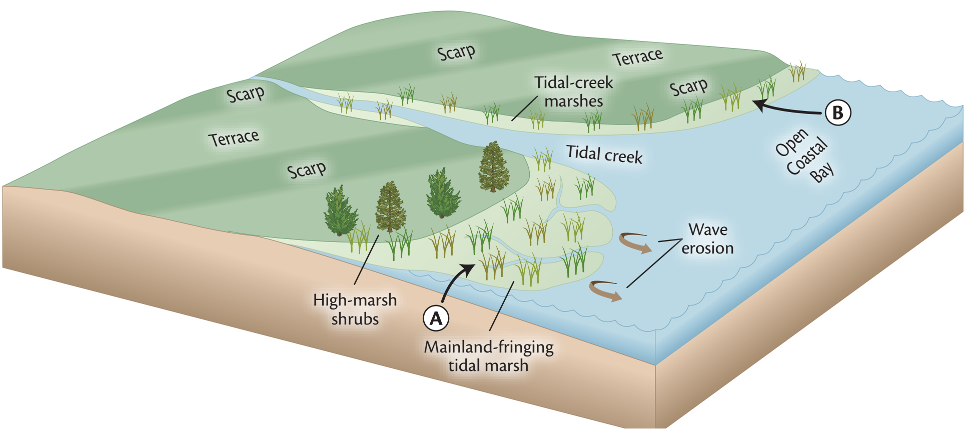
A renourished beach is an effective buffer for storm waves and will protect buildings. However, this approach does not allow for accretion, or increase in elevation of the island surface, by layers of overwash sand. If the rate of sea level rise increases over the next 50 to 100 years, proportionally more new sand will be needed to maintain the beach and position of the island. Even so, the low-lying interior of the island will be increasingly susceptible to storm flooding from both the ocean and bay sides.

During periods of accelerated sea level rise, the natural tendency of wave-dominated, microtidal barrier islands is to be breached by shallow inlets and to have extensive washover, which flattens the dunes, moves sand across the island into the back-barrier bay, and moves the island itself rapidly landward. These are the likely responses of Assateague Island and other undeveloped sections such as Delaware Seashore State Park.

Rapid sea level rise also will affect the structure and function of the Coastal Bays. The tidal marshes that cover intertidal overwash flats and flood-tidal deltas on the back barrier, low platforms on the mainland side of the Coastal Bays, and margins of tidal creeks, will be at risk of drowning. Currently, these marshes keep pace with sea level rise by trapping sediment particles and raising the marsh surface. However, if accelerated sea level rise outpaces the supply of fine suspended sediment, the marsh surface will drop below mean low tide, and the marsh grasses will drown, creating large open areas, or pannes, in the marsh. Additionally, the outer edges of the marsh will be attacked more vigorously by waves and will erode quickly. Marsh islands in the Coastal Bays, which are important nesting areas for waterbirds, are likely to vanish.

The setting of a tidal marsh is critical for its response to sea level rise and its likelihood of surviving. As the rising sea floods the area, new tidal wetlands naturally form on the landward margin of

Sea level rise and marshes



As sea level rises, mainland-fringing marshes that are flanked on the landward side by a low-lying land surface (site A) can expand and maintain marsh area. Marshes adjacent to a steep landward slope, such as a scarp (site B), are likely to be eroded and destroyed. As tidal influence advances upstream in creeks, new marsh area will be created along the margins of the creek.

existing wetlands (for more information on wetlands, see Chapter 14—*Habitats of the Coastal Bays and Watershed*). A flat, low-lying land surface adjacent to a mainland-fringing marsh allows expansion to maintain the area of the marsh as the seaward edge erodes. In contrast, a marsh that abuts a steep slope to the upland, such as a scarp, can not expand landward, and ultimately will be destroyed by the rising sea. Even as sea level rises, new marsh area may be created in protected intertidal areas behind forming, prograding spits, and along the margins of creeks draining the upland as tidal influence migrates upstream.

Existing tidal marshes on back-barrier flats, overwash fans, and flood-tidal deltas behind the barrier islands are gradually buried by new overwash deposits. However, as the barrier-island system retreats landward with continued sea level rise and old marshes are buried, new intertidal flats are created and can be colonized by marsh grasses.

One potential benefit to the Coastal Bays may result from rapid sea level rise. If Assateague Island is breached at one or more sites to create stable tidal inlets, the flushing of the Coastal Bays with seawater will increase dramatically. The most likely locations for new inlets are the low areas of Assateague Island that were previous inlets, such as Green Run, Fox Hills Level, and Little Level. Engineers are likely to fill and close off any breaches that might occur on Fenwick Island (Ocean City) before a permanent inlet could be established.

A new, stable inlet to Chincoteague Bay would have profound effects on the entire Coastal Bays system, and it probably would become more like the lagoons behind the Virginia barrier islands. The tidal range in the Coastal Bays would increase from 12 cm (4.7 in) to 1.2–1.5 m (4–5 ft), equivalent to the open ocean. Enhanced exchange with the ocean would partially alleviate nutrient enrichment and invigorate marsh growth both by increasing the amount of suspended sediment transported to the marsh and by flushing marsh soils of toxic, chemically reduced compounds, such as hydrogen sulfide. The salinity of the Coastal Bays would rise, and large areas would become intertidal flats.

Coastal systems are dynamic and adapt to changing physical conditions. Our perception of the effects of sea level rise and coastal retreat depends largely upon whether the barrier islands are relatively natural parks or highly developed resort areas. Management of each area will differ, but decision-makers need accurate predictions of coastal responses with the likelihood of increased frequency and intensity of coastal storms and accelerated sea level rise associated with global warming.

TIDAL INLETS & COASTAL ENGINEERING

Inlets migrate naturally

The creation of the Ocean City Inlet by the hurricane of 1933, separating Fenwick Island from Assateague Island to the south, was not a unique event. Inlets have formed, filled in, re-formed, and migrated throughout time. Many inlets have existed previously along Assateague and Fenwick Islands.

John C. Dietz, Central Photographic Facility.

This photo, taken in January 1963, shows the inlet formed during the 1962 storm. This view is looking approximately southwest, with the Ocean City airport in the background.

Inlets have been a significant source of sediments to the Coastal Bays in the past. When the water velocity of an incoming tide decreases, sediment falls out of the water column and is deposited on the Coastal Bays side of the inlet, creating a flood-tidal delta. In areas where previous inlets have closed, their flood-tidal deltas remain as islands in the bays which eventually erode and become a source of sediments.

Engineering projects influence sand movement & inlet migration

In 1934, the U.S. Army Corps of Engineers stabilized Ocean City Inlet by constructing two rock jetties, one on the southern end of Fenwick Island and the other on the northern end of Assateague Island. This construction had profound consequences for sediment dynamics in the Coastal Bays. The jetties disrupted the southward longshore transport of sand, which once traveled the length of the Assateague peninsula.

In addition to the longshore transport impeded by the jetties, the inlet itself became a sink for sand that would otherwise reach Assateague Island. The twice-daily flooding tides transport sand through the inlet and deposit it as a flood-tidal delta and tidal shoals in Isle of Wight and Sinepuxent Bays. Some of these sand bodies are seen clearly just north of the Route 50 bridge in the sequence of aerial photographs on the facing page. Similarly, strong ebbing tides carry sand seaward out of the inlet and deposit it as a shallow ebb-tidal delta, or ebb shoal, that may extend nearly 2 km (1.2 mi) into the ocean from the inlet mouth. The outer edge of the shoal is marked by breaking waves that can be seen on all but the calmest days.

The U.S. Army Corps of Engineers documented evolution of the ebb shoal to -13 m (-43 ft) National Geodetic Vertical Datum from 1933 to 1995.[46] The ebb shoal grew rapidly in area from 1933 to 1962, but was relatively stable from 1962 through 1995. The volume of the ebb shoal

U.S. Army Corps of Engineers.

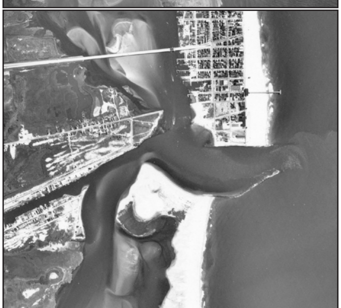
In this photo, both the ebb-tidal shoal (foreground) and the flood-tidal shoal (upper right) are visible. Part of the flood-tidal shoal, north of the Route 50 bridge, is actually an island, Skimmer Island—an important shorebird and waterbird habitat. Also visible is the ebb-tidal plume coming out of the inlet.

The changing dynamics of the Ocean City Inlet

September 18, 1933, a month after the 1933 storm breached the island to create the Ocean City Inlet. The ebb-tidal shoal is already visible as breaking waves offshore.

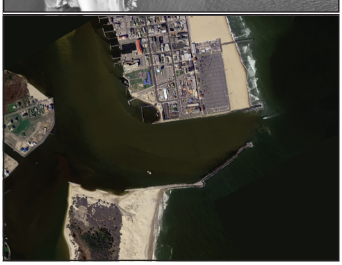
October 9, 1934, approximately 13 months after the creation of the inlet. Here, the U.S. Army Corps of Engineers has begun to stabilize the inlet by constructing jetties. The inlet is substantially wider than when it was created, and the ebb- and flood-tidal shoals are becoming more extensive.

December 6, 1935. Stabilization of the inlet has changed the patterns of sediment transport, with sand accumulating on the northern side of the Ocean City jetty, and on the southern, landward side of the Assateague Island jetty.



June 8, 1952. The northern jetty is completely impounded with sand transported from the north, while the northern Assateague Island beach has retreated more than 115 m (375 ft). Extensive flood-tidal shoals are present in Isle of Wight Bay (west of Ocean City) and in Sinepuxent Bay (west of Assateague Island).

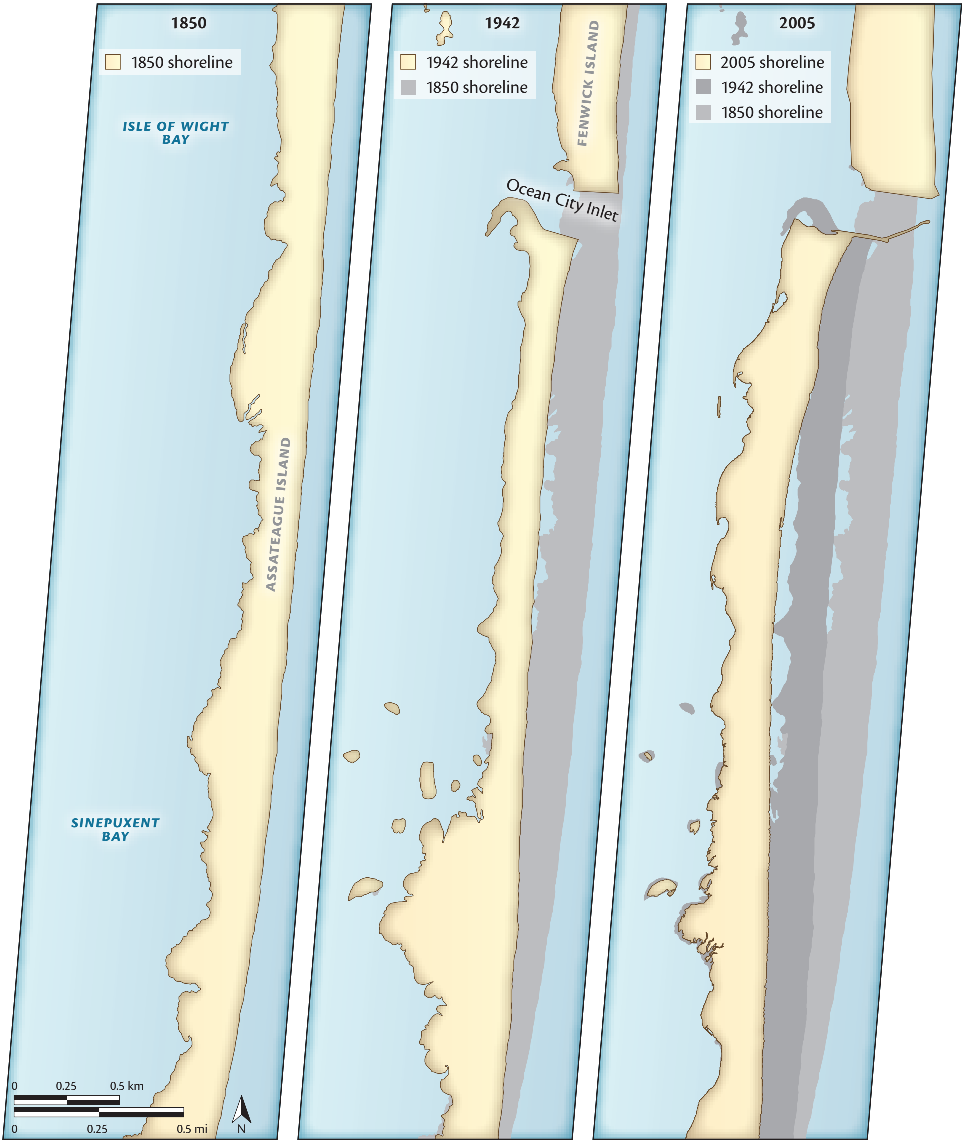
April 28, 1962, approximately one month after the Ash Wednesday nor'easter storm. The northern end of Assateague Island was submerged during the storm, and subsequently detached from the southern jetty. Much of the sand that eroded during the storm was washed into the channel behind the island.



April 16, 2004. This recent photo shows the current location of the Ocean City Inlet and islands.

U.S. Army Corps of Engineers and Maryland Department of Natural Resources.

The evolution of Fenwick and Assateague Islands, 1850–2005



Andrew Roach for Assateague Island National Seashore.

This series of maps shows the shoreline change of Fenwick and Assateague Islands since 1850. After the Ocean City Inlet opened in 1933 and was stabilized the following year, the northern end of Assateague Island has been migrating steadily landward.

increased continuously from 1933 through 1995, although the most rapid rate was immediately after stabilization of the inlet.

The ebb shoal is still growing, and the ultimate equilibrium volume could exceed 3,000,000 m^3 (4,000,000 yd^3) of additional sand beyond its 1995 volume—this volume possibly could be reached by 2040.[22] The U.S. Army Corps of Engineers is currently reassessing growth trends of the ebb shoal. Recent tightening of the south jetty now shunts sand farther offshore than previously, which effectively increases the volume needed for the ebb

Date[1]	Volume		Area		Volume increase	
	Millions of m^3	Millions of yd^3	Millions of m^2	Acres	Thousands of m^3/year	Thousands of yd^3/year
June 1933	0	0	0	0	0	0
March 1937	1.3	1.7	0.822	203	317	415
May 1962	4.4	5.7	3.339	825	123	161
January 1978	8.9	11.7	3.671	907	290	379
October 1995	10.3	13.5	3.638	899	79	103

u.s. Army Corps of Engineers.

Ebb shoal volume, area, and growth rate since stabilization of the Ocean City Inlet in 1933. Volume and area are to a depth of 13 m (43 ft).

1. The data presented for January 1978 are derived from surveys conducted in August 1977 and October 1978. The data presented for October 1995 are derived from surveys conducted in July, October, and December of that year.

shoal to reach dynamic equilibrium with the physical environment.

Barrier islands naturally migrate landward through overwash processes, but sand deprivation on the northern end of Assateague Island has accelerated beach erosion. In the area of the jetty, the island has retreated its entire width (about 500 m or 0.3 mi) since the inlet was stabilized.

This sediment starvation threatens Assateague Island with accelerated erosion and unnaturally high overwash potential due to lowered elevations. To address the cause of these threats, the u.s. Army Corps of Engineers and the National Park Service joined with the Town of Ocean City, Worcester County, and the Maryland Department of Natural Resources to create the Long-Term Sand Management Project, which restores the southward transport of sand from Ocean City towards Assateague Island.

Wave energy, tidal currents, and sand transport in the vicinity of the Ocean City Inlet have been studied extensively by the u.s. Army Corps of Engineers,[46] Maryland Geological Survey, and others, and the inlet is among the best studied in the world. Twice each year, 72,000 m^3 (94,000 yd^3) of sand are moved from the ebb- and flood-tidal deltas around the Ocean City Inlet and deposited in the surf zone of Assateague Island, approximately 2.5–5 km (1.5–3 mi) south of the inlet. Natural processes of waves and longshore transport then deposit this sand in the surf zone and on the beach.

This project and another long-term project called the Atlantic Coast of Maryland Shoreline Protection Project, funded by federal, state, county, and city governments, also replenish the Ocean City beach and dunes with sand. Beach replenishment at Ocean City, in addition to the development of permanent structures on the island, means that the natural landward migration of Fenwick Island has been all but halted.

Hydrodynamic models developed by the u.s. Army Corps of Engineers show that there is a nodal point off Assateague Island, approximately 4 km (2.5 mi) south

Project year	Millions of m^3	Millions of yd^3
1988	1.76	2.30
1990	1.68	2.20
1991	1.24	1.62
1992	1.22	1.59
1994	0.96	1.25
1998	0.99	1.29
2002	0.57	0.74
2006	0.71	0.93
Total volume 1988–2006	**9.11**	**11.92**

u.s. Army Corps of Engineers.

Volume of beach nourishment sand added to Ocean City, Maryland, shoreline.[48]

of the Ocean City Inlet. North of this point, sediment has a net northward transport toward the jetty. South of this point, net longshore transport moves sand southward.

This nodal point is the reason why the long-term sand management project deposits sand about 2.5–5 km (1.5–3 mi) south of the inlet, instead of starting at the northernmost point of Assateague Island. This difference is due to the morphology of the ebb-tidal delta or shoal, which curves around the Ocean City Inlet and, at its southernmost point, is nearly perpendicular to Assateague Island.

Similar to the sheltering effect of Cape May Shoals and Hen and Chickens Shoal at the mouth of Delaware Bay, but on a smaller scale, the ebb-tidal shoal partially protects the Ocean City Inlet from wave attack. Waves approaching from the northeast are refracted around the shoal, changing the direction at which they hit the Assateague Island shoreline south of the shoal. Because of the wave refraction, resulting longshore transport is to the north immediately south of the inlet.

The physical setting and dynamic processes of the Coastal Bays provide the foundation for a complex and productive ecosystem. The interaction of the bays with the mainland watershed and the coastal ocean is mediated largely by fluxes of water, dissolved compounds, and suspended particles transported by surface water, groundwater, and tidal flow through inlets. Against the background of daily to seasonal weather cycles, a few major storms in a 50- to 100-year period exert a powerful influence by reshaping and restructuring the barrier islands and the Coastal Bays, and altering their function for decades to come. The bays inherit much of their character from their geologic legacy from preceding sea level highstands and lowstands of the late Pleistocene, and have continued to evolve over the past 5,000 years as sea level has risen slowly but continuously. Stresses to the Coastal Bays system, and management challenges for the coming century, are related largely to the increasing pace of development in the coastal zone, and the prospect of accelerated sea level rise driven by global warming.

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REFERENCES

1. Amrhein, J.L. Jr. 1986. *The Ancient Seacoast of Maryland including part of Delaware and Virginia: 24-inch by 36-inch chart (map)*. Ryan and Black Publishers, Salisbury, Maryland.
2. Bachman, L., & J. Wilson. 1984. *The Columbia Aquifer of the Eastern Shore of Maryland*. Maryland Geological Survey Report of Investigations 40, Annapolis, Maryland.
3. Boesch, D.F., & J. Greer (eds). 2003. *Chesapeake Futures*. Chesapeake Bay Program Scientific and Technical Advisory Committee, Annapolis, Maryland.
4. Bratton, J.F., J.K. Böhlke, F.T. Manheim, & D.E. Krantz. 2004. Geochemistry of pore fluids and discharge of submarine ground water in Delmarva Peninsula coastal bays. *Ground Water* 42: 1021–1034.
5. Bratton, J.F., J.K. Böhlke, D.E. Krantz, & C.R. Tobias. In press 2008. Flow of groundwater beneath a back-barrier lagoon: Geometry and geochemistry of the subterranean estuary at Chincoteague Bay, Maryland, U.S.A. *Marine Chemistry*.
6. Bruun, P., 1962. Sea level rise as a cause of shore erosion. *Journal of Waterways and Harbors Division, American Society of Civil Engineers Proceedings* 88: 117–130.
7. Cooke, C.W. 1931. Seven coastal terraces in the southeastern states. *The Journal of the Washington Academy of Science* 21: 503–513.

8. Dalrymple, R.A., & D.W. Mann. 1985. *A Coastal Engineering Assessment of Fenwick Island, Delaware.* University of Delaware, Department of Civil Engineering, Technical Report. # CE-54, Oct.1985, Newark, Delaware.
9. Demarest, J.M. 1981. *Genesis and Preservation of Quaternary Paralic Deposits on Delmarva Peninsula*. PhD dissertation, University of Delaware, Newark, Delaware.
10. Demarest, J.M., & S.P. Leatherman. 1985. Mainland influence on coastal transgressions: Delmarva Peninsula. *Marine Geology* 63: 19–33.
11. Dillow, J.J.A., & E.A. Greene. 1999. *Ground-water Discharge and Nitrate Loadings to the Coastal Bays of Maryland*. U.S. Geological Survey Water-Resources Investigations Report 99-4167, Maryland.
12. Dolan, R., H. Lins, & J. Stewart. 1980. *Geographical Analysis of Fenwick Island, Maryland, a Middle Atlantic Coast Barrier Island.* U.S. Geological Survey Professional Paper 1177-A, Washington, D.C.
13. Douglas, B.C., M.S. Kearney, & S.P. Leatherman (eds). 2001. *Sea Level Rise: History and Consequences*. Academic Press, San Diego, California.
14. Douglas, B.C., & W.R. Peltier. 2002. The puzzle of global sea-level rise. *Physics Today* 3: 35–40.
15. Fisher, J.J. 1967. Origin of barrier island shorelines: Middle Atlantic States. *Geological Society of America Special Paper* 115: 66–67.
16. Hall, S.Z. 2005. *Hydrodynamics of Freshwater Ponds on a Siliciclastic Barrier Island, Assateague Island National Seashore, Maryland*. Master of Science thesis, University of Maryland, College Park, Maryland.
17. Hayes, M.O. 1979. Barrier island morphology as a function of tidal and wave regime. *In:* Leatherman, S.P. (ed). *Barrier Islands from the Gulf of St Lawrence to the Gulf of Mexico*. Academic Press, New York, New York.
18. Intergovernmental Panel on Climate Change (IPCC). 2001. *Climate Change 2001: The Scientific Basis*. Contribution of Working Group I to the Third Assessment Report of the United Nations Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, United Kingdom and New York, U.S.A.
19. Intergovernmental Panel on Climate Change (IPCC). 2007. *Climate Change 2007: The Physical Science Basis*. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, United Kingdom and New York, U.S.A.
20. Krantz, D.E., F.T. Manheim, J.F. Bratton, & D.J. Phelan. 2004. Hydrogeologic setting and ground-water flow beneath a section of Indian River Bay, Delaware. *Ground Water* 42: 1035–1051.
21. Krantz, D.E., D.R. Levin, & G. Wikel, unpublished data from a seismic survey conducted in May 2005.
22. Kraus, N.C. 2000. Reservoir model of ebb-tidal shoal evolution and sand bypassing. *Journal of Waterway, Port, Coastal, and Ocean Engineering* 126: 305–313.
23. Lung, W.S., 1994. *Water Quality Modeling of the St. Martin River, Assawoman, and Isle of Wight Bays*. Final report to Maryland Department of the Environment.
24. Mackintosh, B. 1982. *Assateague Island National Seashore—An Administrative History*. U.S. Department of the Interior, National Park Service, Washington D.C. Available online: *http://www.nps.gov/asis/parkmgmt/index.htm*
25. Manheim, F.T., D.E. Krantz, & J.F. Bratton. 2004. Investigations of submarine ground-water discharge in Delmarva coastal bays by horizontal resistivity surveying and ancillary techniques. *Ground Water* 42: 1052–1068.
26. Martinson, D.G., J.D. Hays, J. Imbrie, T.C. Moore Jr., N.G. Pisias, & N.J. Shackleton. 1987. Age dating and the orbital theory of the ice ages; development of a high-resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
27. McBride, R.A. 1999. Spatial and temporal distribution of historical and active tidal inlets: Delmarva Peninsula and New Jersey, U.S.A. *In:* Kraus, N.C., & W.G. McDougal (eds). *Coastal Sediments '99: Proceedings of the 4th International Symposium on Coastal Engineering and Science of Coastal Sediment Processes.* American Society of Civil Engineers, Reston, Virginia.
28. Miller, L., & B.C. Douglas. 2004. Mass and volume contributions to 20th century global sea level rise. *Nature* 248: 407–409.
29. Mixon, R.B. 1985. *Stratigraphic and Geomorphic Framework of Uppermost Cenozoic Deposits in the Southern Delmarva Peninsula, Virginia and Maryland.* U.S. Geological Survey Professional Paper 1067-G, Washington, D.C.
30. Moore, W.S. 1999. The subterranean estuary: A reaction zone of ground water and sea water. *Marine Chemistry* 65: 111–125.
31. National Research Council, Geophysics Study Committee, Commission on Physical Sciences, Mathematics and Resources. 1990. *Sea Level Change*. National Academy Press, Washington, D.C.
32. NOAA National Data Buoy Center. 2006. Station 44009 off Delaware Bay. Historical data and climatic summaries. *http://www.ndbc.noaa.gov*
33. NOAA National Ocean Service, Sea Levels Online. *http://tidesandcurrents.noaa.gov/sltrends/sltrends.shtml*
34. Oertel, G.F., & J.C. Kraft. 1994. New Jersey and Delmarva barrier islands. *In:* Davis, R.A., Jr. (ed). *Geology of Holocene Barrier Island Systems*. Springer-Verlag, Berlin, Germany.
35. Owens, J.P., & C.S. Denny. 1978. *Geologic Map of Worcester County.* Maryland Geological Survey, scale 1:62500.
36. Pendleton, E.A., S.J. William, & E.R. Thieler. 2004. *Coastal Vulnerability Assessment of Assateague Island National Seashore (ASIS) to Sea-Level Rise.* U.S. Geological Survey Open-File Report 2004-1020.
37. Pritchard, D.W. 1960. Salt balance and exchange rate for Chincoteague Bay. *Chesapeake Science* 1: 48–57.
38. Ramsey, K.W., & S.J. Baxter. 1996. *Radiocarbon Dates from Delaware: A Compilation*. Delaware Geological Survey Report of Investigations no.54, University of Delaware, Newark, Delaware.
39. Ramsey, K.W., D.J. Leathers, D.V. Wells, & J.H. Talley. 1998. *Summary Report: The Coastal Storms of January 27–29 and February 4–6, 1998, Delaware and Maryland*. Delaware Geological Survey, Open File Report 40, University of Delaware, Newark, Delaware.
40. Shedlock, R.J., & D.W. Bolton. 2006. *Sustainability of the Ground-water Resources of the Atlantic Coastal Plain of Maryland*. U.S. Geological Survey Fact Sheet 2006-3009.
41. Stauble, D.K., A.W. Garcia, N.C. Kraus, W.G. Grosskopf, & G.P. Bass. 1993. *Beach Nourishment Project Response and Design Evaluation: Ocean City, Maryland, Report 1, 1988–1992.* Technical Report CERC-93-13, USAE Waterways Experiment Station, Vicksburg, Mississippi 39180-6199.
42. Titus, J.G., S.P. Leatherman, C.H. Everts, D.L. Kriebel, & R.G. Dean. 1985. *Potential Impacts of Sea Level Rise on the Beach at Ocean City, Maryland.* Environmental Protection Agency, Washington, D.C.
43. Toscano, M.A., R.T. Kerhin, L.L. York, T.M. Cronin, & S.J. Williams. 1989. *Quaternary Stratigraphy of the Inner*

Continental Shelf of Maryland. Maryland Geological Survey Report of Investigation 50, Baltimore, Maryland.
44. Toscano, M.A., & L.L. York. 1992. Quaternary stratigraphy and sea-level history of the U.S. Middle Atlantic Coastal Plain. *Quaternary Science Reviews* 11: 301–328.
45. U.S. Army Corps of Engineers. 1962. *The March 1962 Storm Along the Coast of Maryland: A Report of District Activities During and Immediately Following the Storm*. Baltimore District, Baltimore, Maryland.
46. U.S. Army Corps of Engineers. 1998. *Ocean City, Maryland, and Vicinity Water Resources Study. Final Integrated Feasibility Report and Environmental Impact Statement, Appendix D, Restoration of Assateague Island*. Baltimore, Maryland.
47. U.S. Army Corps of Engineers. 2006. Climatic summary and extremal analysis for Station MD002 Ocean City, Maryland. USACE Engineering Research and Development Center. *http://sandbar.wes.army.mil*
48. U.S. Army Corps of Engineers. 2007. *Atlantic Coast of Maryland Shoreline Protection Project, Supplemental Environmental Impact Statement, General Reevaluation Study: Borrow Sources for 2010–2044*. May 2007 Draft. U.S. Army Corps of Engineers, Baltimore, Maryland.
49. Wazniak, C., L. Karrh, T. Parham, M. Naylor, M. Hall, T. Carruthers, & R. Orth. 2004. Section 6: Habitat condition in the Maryland Coastal Bays. *In:* Wazniak, C.E., & M.R. Hall (eds). *Maryland's Coastal Bays Ecosystem Health Assessment 2004*. DNR-12-1202-0009. Maryland Department of Natural Resources, Tidewater Ecosystem Assessment, Annapolis, Maryland.
50. Wehmiller, J.F., K.R. Simmons, H. Cheng, R.L. Edwards, J. Martin-McNaughton, L.L. York, D.E. Krantz, & C.-C. Shen. 2004. Uranium-series coral ages from the U.S. Atlantic Coastal Plain—the '80 ka problem' revisited. *Quaternary International* 120: 3–14.
51. Wells, D.V. 1994. *Non-energy Resources and Shallow Geological Framework of the Inner Continental Margin off Ocean City, Maryland*. Maryland Geological Survey Coastal and Estuarine Geology Program Open File Report 16, Maryland.
52. Wells, D.V, & R. Conkwright. 1999. *The Maryland Coastal Bays Sediment Mapping Project—Physical and Chemical Characteristics of the Shallow Sediments: Synthesis Report and Atlas*. Maryland Department of Natural Resources, Maryland Geological Survey, Coastal and Estuarine Geology Program File Report 99-5, Annapolis, Maryland.
53. Zhang, K., B.C. Douglas, & S.P. Leatherman. 2004. Global warming and coastal erosion. *Climatic Change* 64: 41–58.