Geologic history of Chesapeake Bay region – Tertiary to present

We next pick up with a brief history of the late Tertiary development of the Chesapeake Bay region. Check your geologic time scale for approximate dates.

The main point to be made here is that there were alternating periods of rising and falling sea level throughout the Tertiary. These occurred on much longer time scales than during the Pleistocene, and were not attributable to glacial/interglacial cycles, which really didn't get started until the Pleistocene.

Some of the local changes in sea level may have been attributable to regional uplift (question: how would uplift affect local "relative" sea level? If the land is rising relative to the water, the water level appears to drop relative to the land even if global sea level is not changing…) Others may have been attributable to climate variations, though not to glacial episodes. But what we can see from our curve showing long-term trends in sea level is that, from about 15 million years ago until the beginning of the Pleistocene, each successive high stand of sea level was a bit lower than the one before it. This means that each new high stand did not extend as far inland or as high over the land surface as the one before it. What is the practical implication of this?

Terraces and inset deposits

Consider for a moment that the overall stratigraphy (sedimentary layering) of the Coastal Plain is that of a sedimentary wedge, thickening seaward, with the oldest layers on the bottom and the youngest layers on top. The oldest sediments exposed on land are fluvial sands and gravels of the Chesapeake Potomac group, and these occur in the westernmost part of the coastal plain near the Fall Line, where they just barely overlap the crystalline metamorphic rock of the Piedmont. As you go to the east/southeast, progressively younger layers are exposed at the surface and the older layers are found deeper and deeper below the surface.

These sediments, however, were not deposited in simple wedge-shaped sheets. The fluvial sediments were deposited on the upland surface by rivers that migrated back and forth and dumped their sediment successively in one place and then another. As these rivers reached the shoreline, which was much farther inland than it is today, the remaining sediment reached the ocean in deltas or in coastal embayments, and there it accumulated as coastal or shallow marine sediments. Calvert Cliffs on the west shore of the bay exposes a rich sequence of dominantly marine fossil-rich layers that is one of the great historic fossil-collecting localities of this part of the world.

What would happen during a low stand of sea level? The shoreline would move seaward, and the rivers would cut downward into their valleys because lower sea level causes a steeper gradient and more available potential energy for transporting water and sediment. As they cut down, and as they also migrated laterally, there would be an incised valley that formed a trench within the upland coastal plain sediments. Riverine gravels and sands would be deposited in the channel running down this valley.

As sea level rises again, these valleys begin to fill in. In their downstream portions, the fluvial deposits may be buried by younger marine open-water deposits. A bit further upstream, one might find coastal or estuarine or wetland deposits. A bit further upstream still, out of the direct influence of the tides, one might still find the rivers aggrading their beds and building up their floodplains to bury the incised valleys, because with higher sea level the gradient is lower and there is less potential energy to transport the sediment further downvalley.

But keep in mind that each high stand is a little lower than the one before it. Therefore the estuarine fill or the layer of marine sediment, or the floodplain sediments, if we are a little further upstream, are not going to be high enough to reach the same level as the surrounding upland. The fill from the current high stand is inset at a level below the upland surface.

With the next cycle of sea-level fall and rise, the next sedimentary fill in turn will be inset even lower.

The result is that our "wedge" of unconsolidated coastal plain sediments is dissected by a series of valleys running toward the sea, and each of these valleys contains a staircase of different levels deposited during successive high stands of sea level. These remnant levels are called terraces. If they were made by rivers, they are fluvial or alluvial terraces; if made by estuarine sedimentation or by marine sediment, then they are estuarine or marine terraces.

If you look at the handouts provided on the Blackboard page, you will see three examples, one from the area around Cecil and Harford county, one from a cross-section of the James River estuary, and one from the mouth of the Bay along the Virginia shore south of the James River. Although they represent different locations around the bay region, they show similar features. The different terraces get progressively younger as you approach lower and lower elevations. Only the lowest couple of levels are of Pleistocene age. If you look at the cryptic table with the sea-level curve on the right side, you can probably find the names of the formations associated with some of these terraces and the approximate time periods during which those sediments were deposited. There is some discussion of this in the Pazzaglia, et al. field-trip guidebook posted in the Geologic Background folder.

Note that the rolling upland topography of the western shore of the Bay is considerably higher than the topography of the eastern shore. The eastern shore developed more recently, during periods when sea level was not as high. Large parts of today's lower eastern shore developed during the most recent high stand of sea level, which was probably no more than 10 to 15 feet higher than modern sea level; thus there are very broad areas of very low relief, and large areas that are dominated by wetlands. This lowest estuarine terrace is made up of sediments of the Kent Island, Nassawadox, or Tabb formations. If you look at the map of lands vulnerable to sea-level rise (at the end of this document, also included in the sea-level rise handouts), you’ll see a broad area of land with elevations less than 1.5 m above sea level.

So much for a brief overview of the story on land. The story under the Bay is even more interesting. I won't say as much about it, because the readings provide a clearer picture of this part of our story.

Pleistocene paleochannels

Consider that, especially during the Pleistocene, sea level was much lower than today. Channels were incised deeply into the bottom of what we now think of as the floor of Chesapeake Bay. The main channel system was that of the Susquehanna River. But during the subsequent high stand of sea level, much of that channel would eventually be filled in by sediment. Furthermore, the sediment would show a gradual transition, with rising sea level (visible in the sediments closer to the surface), from fluvial gravels and sands to estuarine fine sands and muds. This is exactly the kind of sequence revealed in the sediments brought up from boreholes that were drilled for the pilings of the Chesapeake Bay bridge.

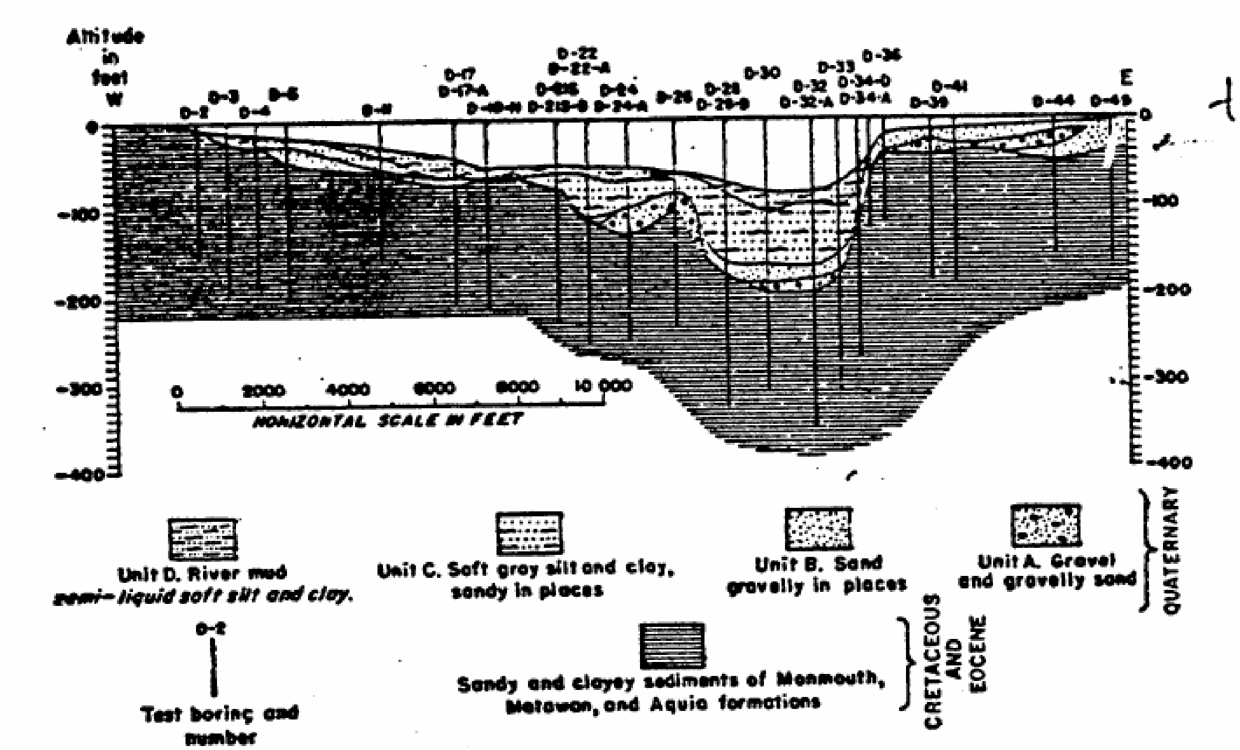


FIGURE 2.—GEOLOGIC CROSS SECTION AT THE SITE OF THE CHESAPEAKE BAY BRIDGE Between Sandy Point and Kent Island (Fig. 1, loc. 1). Modified from Ryan (1953, Pl. 2).

Consider the next low stand. At the seaward end of the Bay, as the water drains out and sea level declines, fluvial incision begins again and cuts a new channel headward, dissecting the landscape that had just been covered by marine and/or estuarine sediment. Does the new channel re-occupy the location of the old one? Or, if the old one was completely obliterated, do we form a new channel in a different place? What evidence do we have to answer this question, and what is the historical sequence of events interpreted from this evidence? This is the primary tale told by one of the papers in the assigned readings, and that paper (Colman and others, 1990) includes several figures illustrating both cross-sections and a plan-view map with three distinct sets of Pleistocene paleochannels representing several successive locations of the Susquehanna River channel during glacial low stands. Another version of this figure is included in the 2006 Master's thesis by Timothy Scott at Old Dominion University. We will have more to say about this in class. We will also discuss the Pleistocene history of climate and sea level in more detail.

One other item needs to be mentioned here. If you read the 1990 paper by Colman and others, the three paleochannels – the Exmore, Eastville, and Cape Charles – are identified as being of latest Pleistocene age (Cape Charles); from the last major glacial period about 150,000 years ago; and from an even earlier interglacial period, either 200,000 or 400,000 years ago (if the older date is correct, then perhaps the same channel might have been reoccupied in the next glacial period). Each of these dates is associated with a low stand of sea level based on the oxygen isotope anomalies that were discussed in class.

However, Timothy Scott's thesis makes an alternative suggestion. He points out that when a major continental glaciation occurs, the isostatic adjustment caused by the tremendous weight of 3-4 km of ice on the continental suface causes the earth's crust to sag and displaces material from the underlying asthenosphere; but as this material flows out from underneath the glacier (much like the water in a waterbed when you sit on it), it causes a bulge out around the margins of the glacier, up to several hundred miles away from the glacial front. So Scott argues that the uplift of the landscape in the local area due to the glacial "forebulge" caused relative sea level to drop even further compared to the rising land surface than would have been the case due to the global drop in sea level alone.

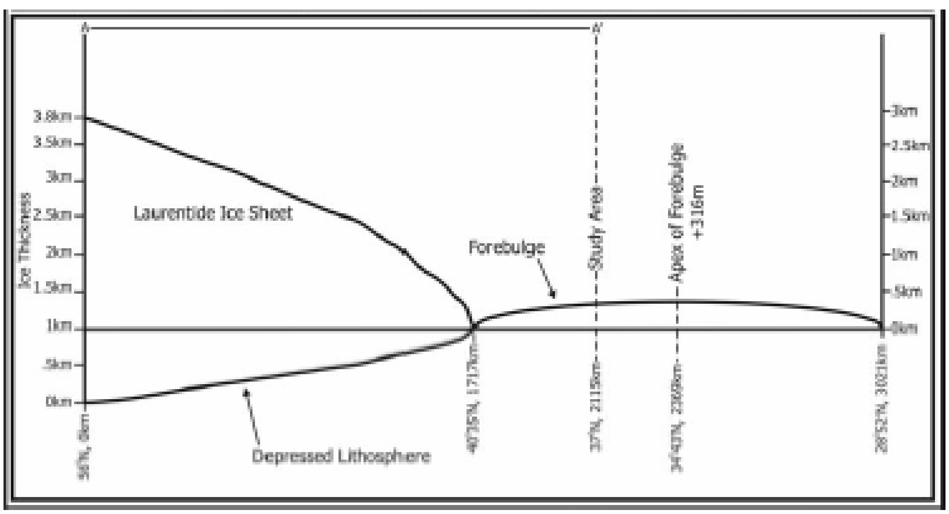


Fig. 49. Laurentide Ice Sheet during the last glacial maximum, modified from Peltier, 1987, and Andrews, 1987, and correlating cross section (vertical exaggeration 2x) to the study area with the associated peripheral bulge. Amount of depressed lithosphere (-1 km) and extent and height of the forebulge were determined in equations 2b-5b. Height and extent of forebulge assumes a complete transfer of mass occurred between the depressed area and the peripheral region.

Furthermore, when the glaciers melt and the displaced material flows back into the area underneath where the glacier had been, the forebulge collapses and the land surface in our area drops, which causes local relative sea level to rise even higher than would have happened from global sea level rise alone. Using other evidence, Scott infers that the Exmore channel formed between 190,000 and 130,000 years ago, during the last major glacial episode; that the Eastville channel formed between 95,000 and 85,000 years ago, in the middle of the last major glacial period, partly as a result of the uplift caused by glacial forebulge; and that the Cape Charles channel formed in the latest Pleistocene, reaching its maximum growth around 18,000 years ago.

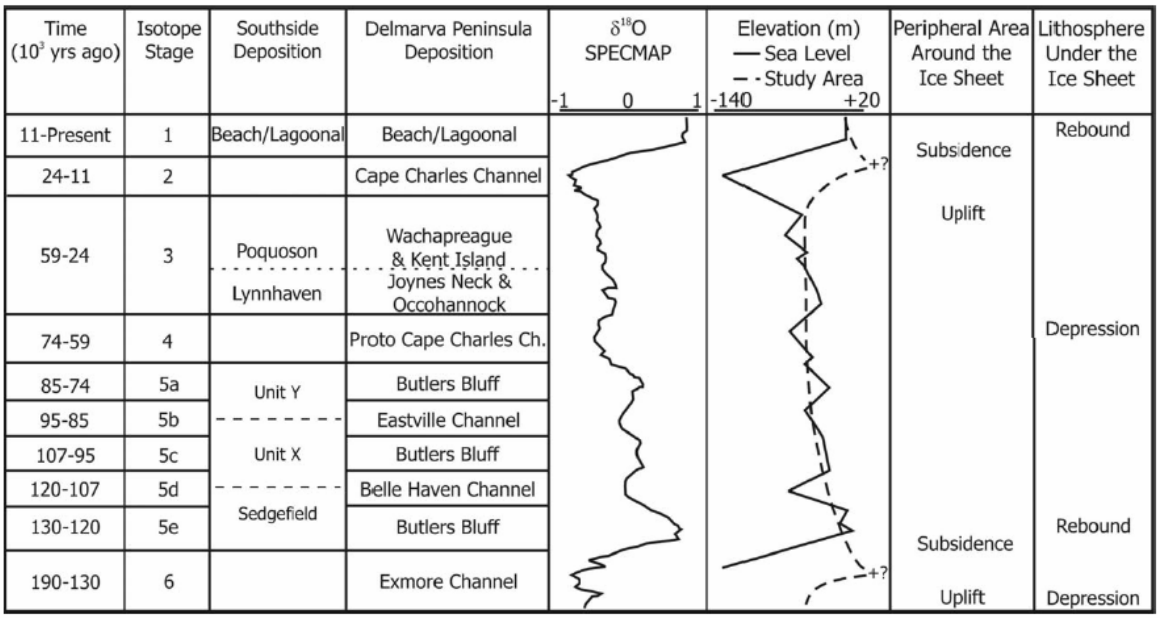


Fig. 59. Revised correlation between the late Pleistocene geologic units on the coastal plain of Virginia with the glacial-eustatic sea-level curve as well as the isostatic adjustments that occurred in the region.

The figure on the next page, also from Scott's thesis, shows stages in the growth of the lower Delmarva peninsula during successive high stands of sea level, followed by renewed channel incision around the south end of the prograding spit during the next glacial period of lower sea level. As the spit grows it buries the older channels, which remain in the subsurface as paleochannels that can be detected using geophysical methods and by drilling boreholes to sample the underlying sediments.

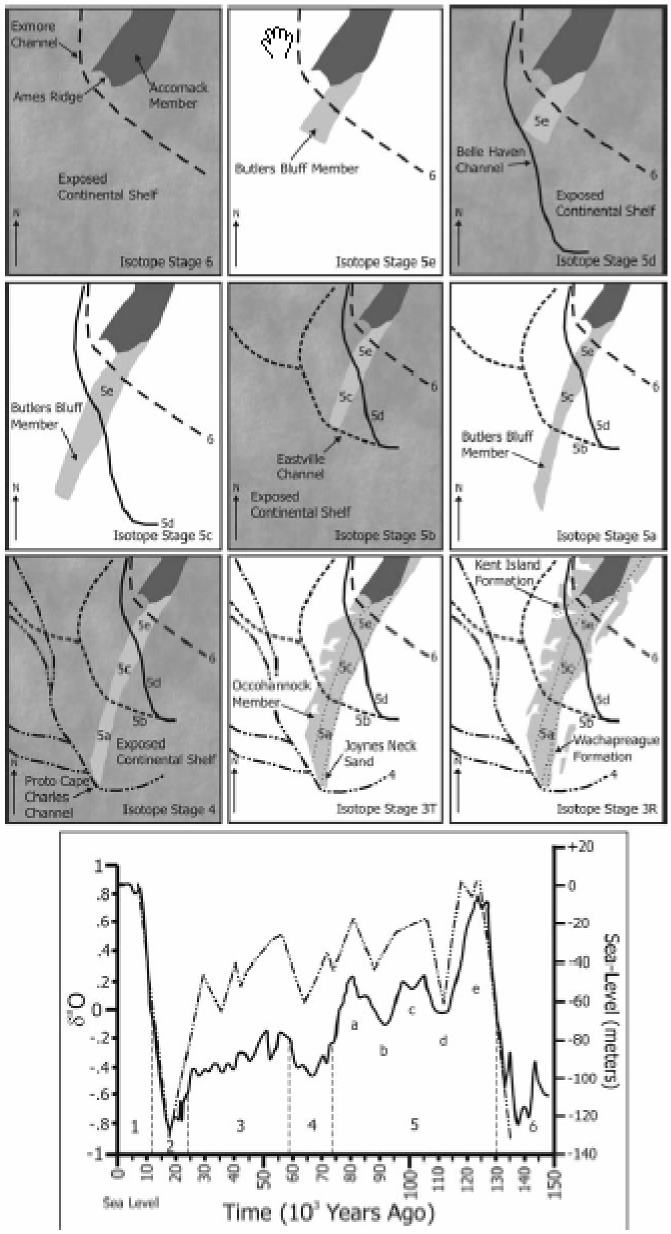
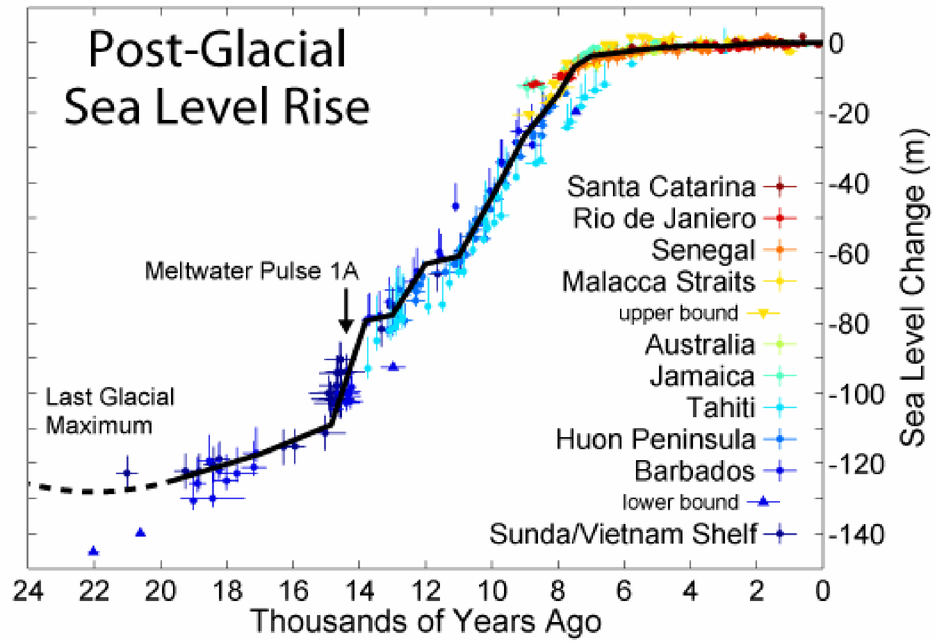
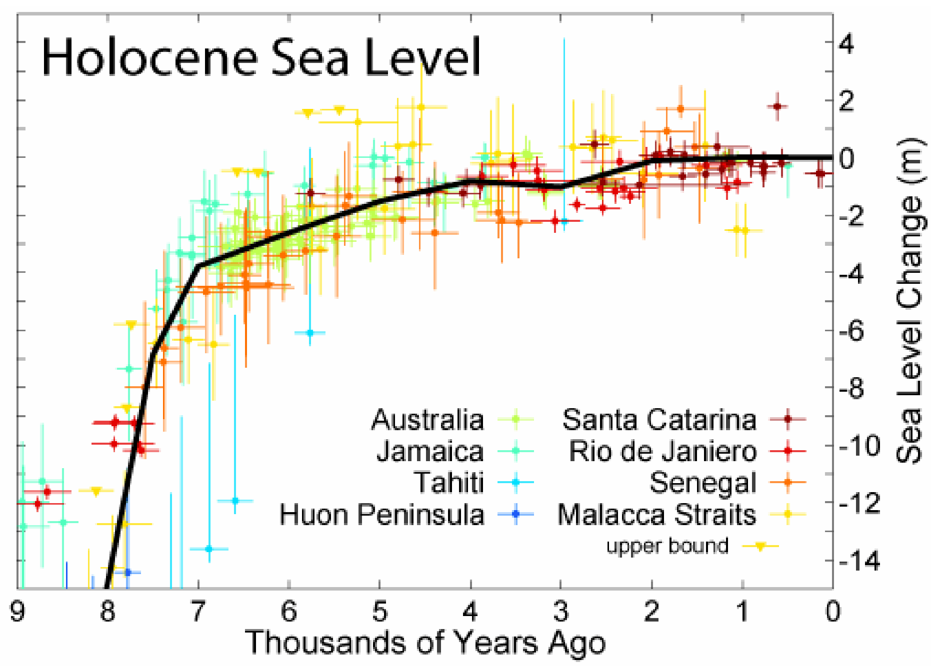


Fig. 40. Proposed deposition sequence of the southern Delmarva Peninsula correlated with the glacial-eustatic sea-level curve modified from Martinson, 1987 and Bradley, 1999. Numbers 3E and 3M represent early stage 3 and middle stage 3 respectively. Paleochannels modified from Parsons *et al.*, 2003, Oertel and Foyle, 1995 and Colman *et al.*, 1990. Isotope Stage 2 lowstand created the "modern" Cape Charles Channel.

Holocene sea-level rise

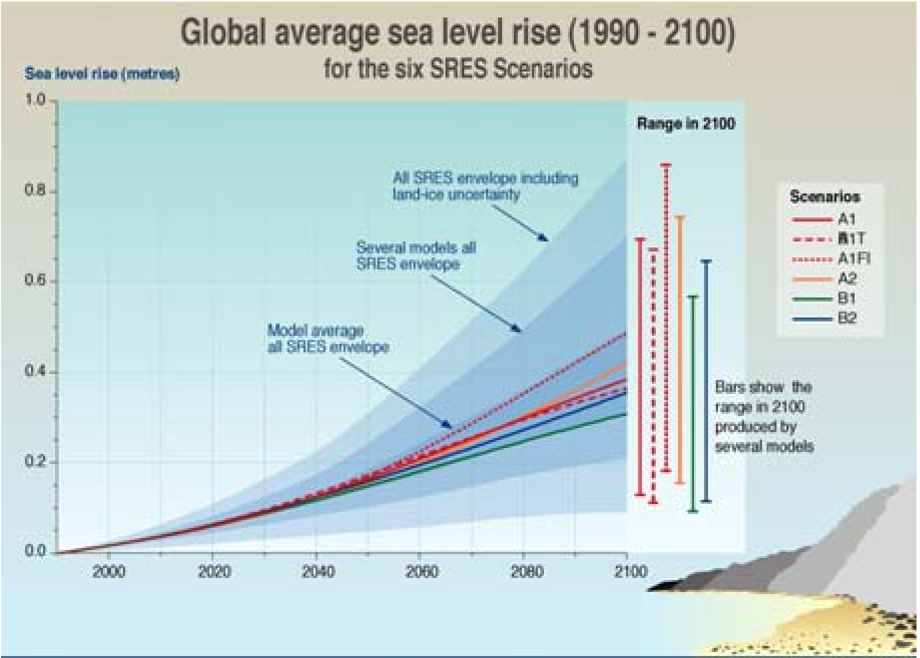
We live in the Holocene, which began about 10,000 years ago. It started with a rapidly rising sea level; the rate of rise gradually tapered off around 7 or 8,000 years ago.





Sea level has continued to rise since then, although at a lower rate. This rate appeared to have slowed down considerably within the last 2000 to 3000 years; yet modern rates in the Chesapeake Bay area seem to show considerable acceleration in recent sea-level rise (not in the figures above, which reflect global data, not local data).

The prospect for continued acceleration as a result of global warming seems quite certain; the only question is how rapid the acceleration will be. We will discuss actual rates in class, as well as some of the differences between global ("eustatic") and local rates of sea-level rise. The IPCC



(Intergovernmental Panel on Climate Change) projections in the figure above are for global sea-level rise and do not account for the effects of local subsidence which could cause higher local rates of sea-level rise. Also, the IPCC projections do not consider the apparent acceleration of ice surges supported by basal meltwater in places like Greenland, which could cause more rapid increases. For recent discussion of the scientific evidence on this point, see the following web site:

http://www.realclimate.org/index.php/archives/2006/03/catastrophic-sea-level-rise-more-evidence-from-the-ice-sheets/

How do we even know anything about the rate of sea-level rise? You need to know (a) the age of an artifact or deposit; and (b) the elevation of sea level at the time it was deposited. The most popular dating tool is radiocarbon, which works primarily on samples containing organic matter. But even if we can date the organic matter, how do we find organic matter that can be clearly identified as representative of sea level?

We'll discuss this in class as well.

Once we have developed a sea-level rise curve, we can start to recognize stages in the progressive growth of Chesapeake Bay under the influence of the advancing ocean as it invades the land. We can estimate, for example, the approximate times when the head of tide reached various locations as the invading sea moved up the Susquehanna River valley. You can find a 2003 paper on the question of when Chesapeake Bay reached its approximate modern extent ("Birth of the modern Chesapeake Bay") on the Course Documents page in the Geologic Background folder. The authors use geochemical evidence to conclude that the transition from fresh water to brackish water in the northern Bay occurred between 8200 and 7400 years ago – right around the time the Holocene sea-level rise decelerated to a much slower rate.

Estuarine geomorphology: coastal modification of a fluvial landscape

The next important question is this: how is the upland fluvial landscape transformed into the estuarine landscape? What are the major processes and what are the telltale signs by which we recognize their influence?

We will explore this in class, and we will also look at some maps or Digital Elevation Models for the visible evidence in today's landscape. I have included a few figures here. (In case you're not aware of it, virtually all of the USGS topographic data for the U.S. are now accessible online.)

What are the major processes?

Before we review them, please note that an estuary is by definition a zone of mixing between freshwater and marine influence. Therefore you will find different things happening in different parts of the system at any given time. The marine portions closer to the mouth will be more dominated by wind, waves, tidal currents, and transport of sediment along the shoreline. They will face a broader stretch of open water and deeper water, so that there is more opportunity for a lot of wave energy to strike the shoreline. By contrast, the headward portions are really tidal rivers; even if they have reversing tidal currents, most of the phenomena we have just discussed will be less influential. At the same time, because of proximity to the head of tide, there will be larger amounts of sediment derived from the river, and deposition rates may be considerably higher. The intermediate zone is just that: it shares some characteristics of both but is not quite dominated by either. We will find that this zonation applies reasonably well to Chesapeake Bay.

But now think about what happens as sea level rises. The head of tide migrates upstream or up-Bay, and therefore the different zones also migrate: a place along the river may start out dominated by the river, but may gradually be inundated by rising sea level, and eventually that same location may be at a point midway down the estuary, where the environmental conditions are very different. Thus, in a sense, the lower portions of the estuary may be in a "later" stage of development than the headward portions. Of course, if sea-level rise slows down or if there is so much sediment coming in that the whole estuary starts to fill in, then the process may actually reverse itself: a delta forms in the estuary and the river reasserts itself by extending its channel back downstream. This really hasn't happened in Chesapeake Bay, except in a few tidewater tributaries that were affected by excessive soil erosion and sedimentation during the period of intensive agriculture. It IS true that the deep channel of the Susquehanna River is buried by Holocene sediment in the part of the Bay north of Annapolis; some people refer to this as an underwater delta but it has not reversed the overall effect of sea-level rise.

Back to the processes:

1) Coastal inundation of the river valley, followed by gradual drowning of the side slopes, eventually inundating even the low divides between adjacent valleys. I refer to the net result of this as "dismembering the drainage network". I will show you some examples from a paper on the paleogeography of Delaware Bay. If you look at nautical charts and maps showing part of Chesapeake Bay, you should be able to reconstruct this same process in your own mind. Dismembering of the drainage network eventually results in a series of tidal creeks all emptying

along a common coastline rather than forming integrated parts of a dendritic drainage system, because the downstream trunk of the system has been completely drowned. Just before the drainage divide is completely drowned, it forms a peninsula between tidal creeks; then, as the low spots on the peninsula are inundated, it separates into islands; finally, even the highest parts are inundated, but the remnant patches of island may be occupied by tidal marsh.

2) Growth of tidal marsh occurs as rising sea level inundates low, relatively flat surfaces. There are two main locations where this may happen: in broad flat terrace areas that are drowned, water may spread out and marshes may grow in a broad swath. This is common on the lower eastern shore. A subsidiary example would occur as marsh starts to grow on the last remnant of a drainage divide as it is inundated, or around the fringes of an island as it is being drowned by rising sea level. Note, however, that in many areas with limited supply of sediment, the marsh cannot keep growing fast enough to stay above water as sea level continues to rise. We are currently losing substantial areas of marsh on the Eastern Shore (including much of Blackwater National Wildlife Refuge) as a result of this phenomenon.

The second example occurs in a tributary valley. As rising sea level inundates the valley, the zone of contact between sea level and the upland stream is one where the stream loses its gravitational energy, drops its sediment, and aggrades its bed. The shallow zone at this location may be rapidly colonized by marsh grasses, and you will find linear patches of marsh grass along this part of the channel system in many small tidal tributaries. With increasing sea level, the marsh may spread over the local floodplain but is still largely confined to the valley floor. Later on, you may find patches of marsh exposed along the open coast by erosion.

3) Shore erosion, largely under the influence of wind-driven waves, tends to straighten shorelines, cutting back protruding headlands and filling in indentations or embayments. An irregular shoreline exposed to enough wave energy will eventually be straightened in this way owing to the effects of wave refraction.

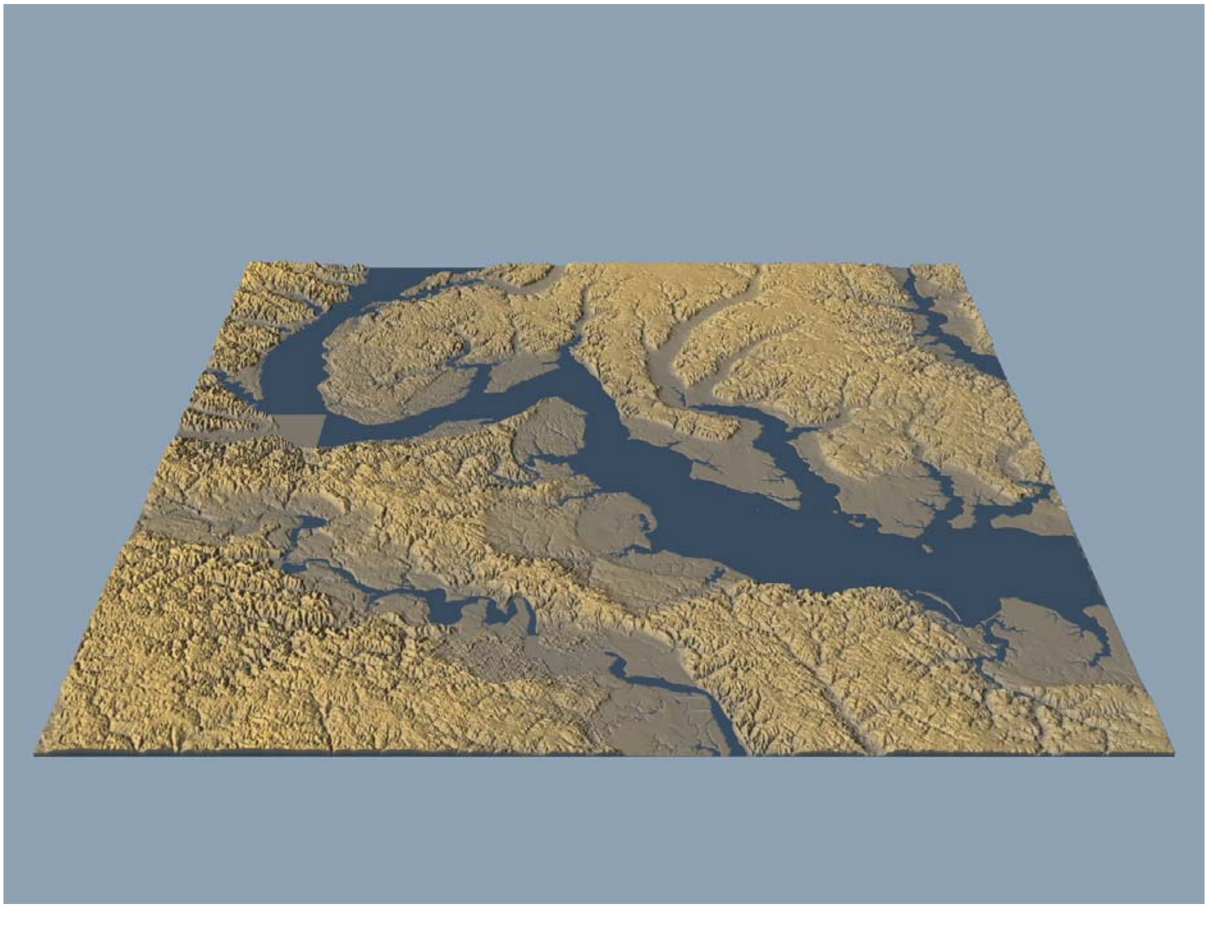
The result is that, after first inundating the fluvial landscape and dismembering the drainage network, we now truncate it: cut back the peninsulas, gradually erasing the fingerlike extensions of the valleys and turning them into short stubs of valleys with flat, relatively straight ends. If we erode far enough back into the valley, we may also expose patches of marsh along the shoreline.

4) Coastal sediment transport occurs primarily under the influence of waves, which rarely hit the shoreline at a perpendicular angle; the result is that there is usually some component of wave activity oriented parallel to the shoreline, and this causes longshore drift. The migration of sand bodies, often derived from winnowing and sorting of the debris left behind by shore erosion, creates spits and bars that may extend out into open water. Bars may grow across the mouths of small- to medium-size tidal creeks, thereby creating a continuous straight shoreline even where indentations were originally present. The inundated portions of smaller tidal creeks may be isolated by these bars to become salt ponds, which eventually will fill in with marsh grass and peat. If the creek is a little bit larger, there may be enough of a tidal current to maintain an inlet through which the tide can enter on the rising portion of the tidal cycle and drain out into the open estuary on the falling portion of the tidal cycle.

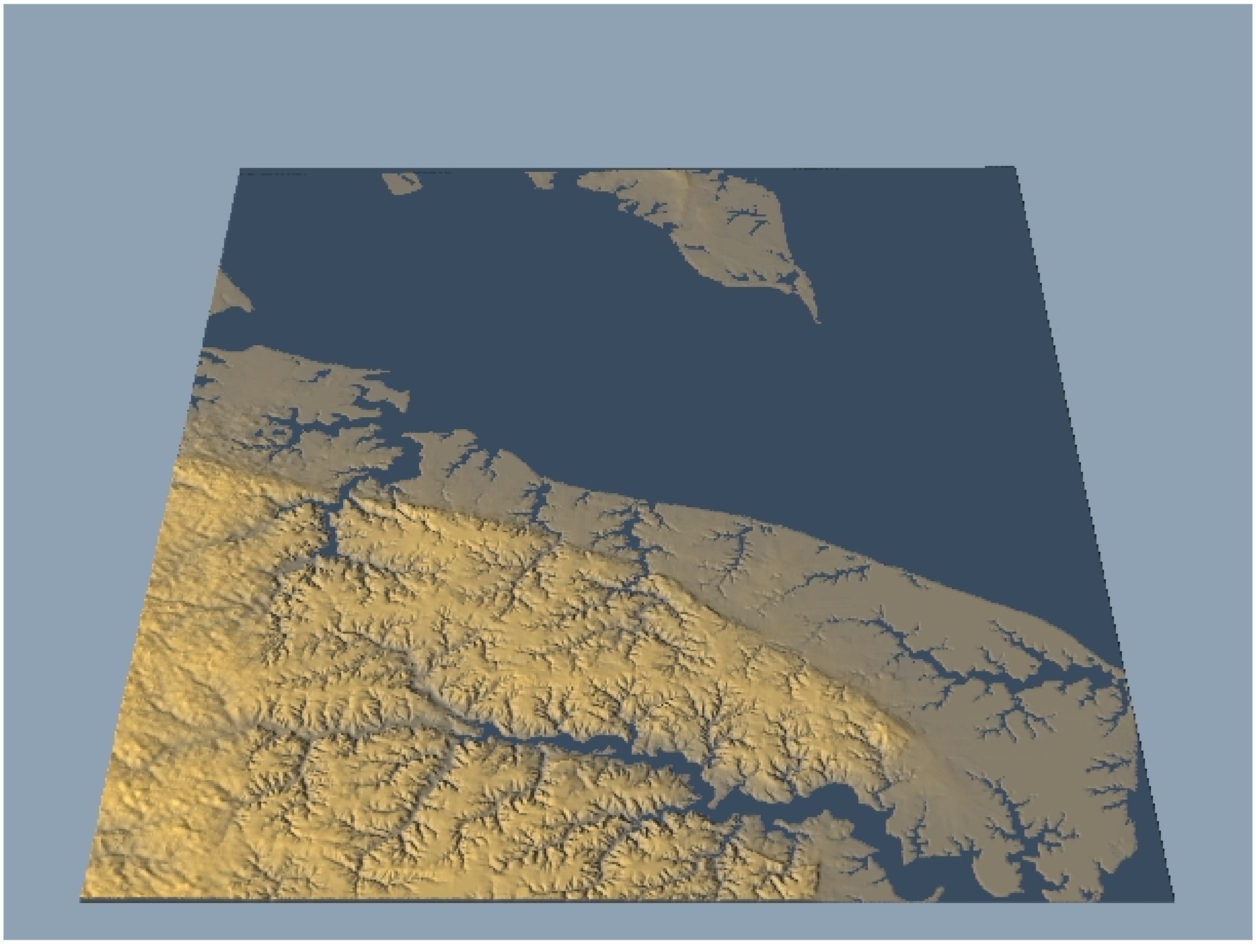
5) Most of what we have discussed up to this point is localized in particular parts of the estuarine system. But deposition of sediment, which will be discussed as a separate topic in a few weeks, also has a profound effect on estuarine geomorphology. Estuaries have a tendency to trap virtually all of the sediment that enters them, for reasons to be discussed later. Therefore they have a limited lifespan before they fill in; the faster sea level rises and the smaller the amount of sediment entering the system, the longer the estuary will survive. In the next couple of centuries, Chesapeake Bay may actually grow slightly larger than it is now. But eventually either sea level will stop rising or sedimentation will overtake other processes and gradually fill in the Bay floor. Most likely this will occur in a headward area that will become a delta, and the delta will gradually grow (or "prograde") downbay until most of the Bay is filled and the estuary becomes a shallow combination of wetland and floodplain. If you go to some of the tributaries of the Bay, such as the Jug Bay portion of the Patuxent River, you can see systems that already look like this.

The end result of this is to create an estuarine fill. The next time sea level falls again, this bay bottom or estuarine fill will be trenched and the remaining portion will be left behind as an estuarine terrace, comparable to the estuarine terrace that makes up most of the lower eastern shore today.

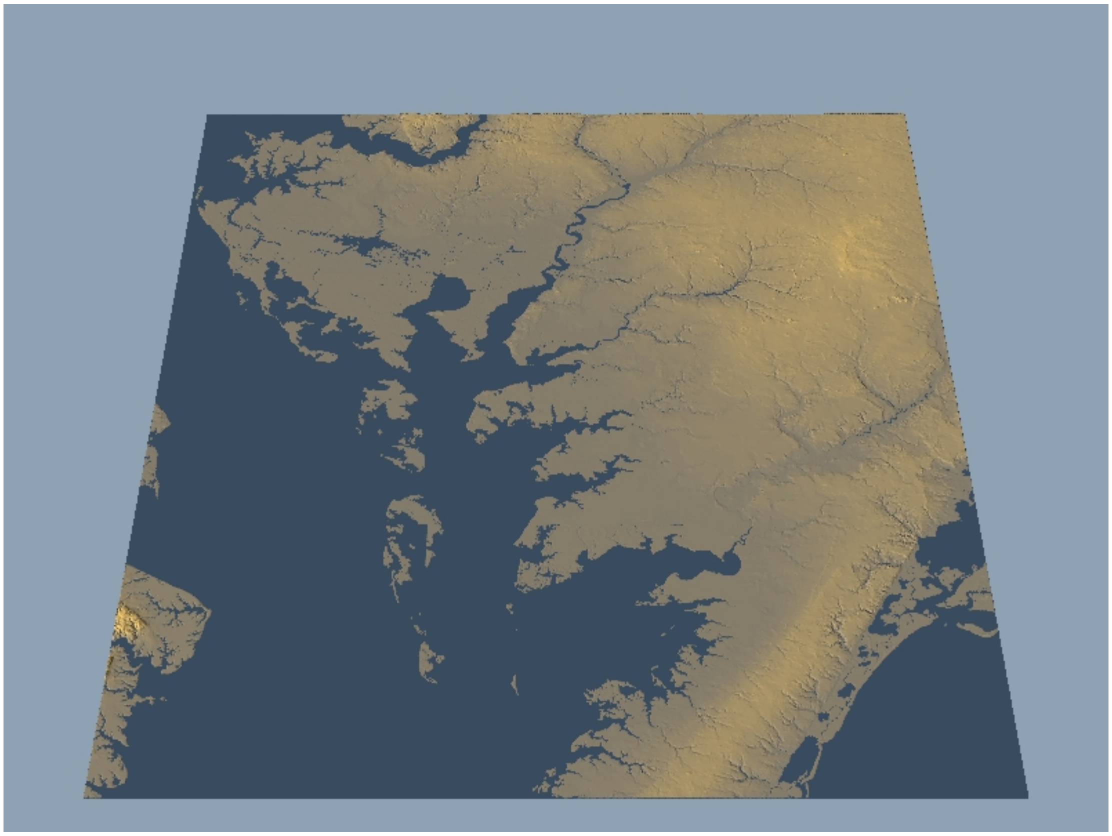
Look at this perspective view of the Potomac estuary, based on a USGS DEM. You can see the different inset surfaces and erosional scarps in the landscape.



If we zoom in on this view of the lower Virginia shore of the Potomac estuary, you can see a wave-cut scarp from the last high stand of sea level; the dendritic drainage pattern of a stream dissecting that older, slightly higher surface; and the truncated remnants of the drainage network that has been dismembered by erosion along the southwestern shore of the Potomac. Erosion together with longshore sediment transport has also straightened and simplified this shoreline by comparison with the highly crenulated and embayed shorelines that you see at more sheltered locations.



If you look at the perspective view of the lower Eastern Shore on the next page, you can also see a chain of islands representing the truncated remnants of a peninsula that was formerly the divide between two stream valleys. These islands are so close to present sea level that they are almost entirely covered by marsh vegetation. The shoreline is not quite so straight as in the example on this page because it has not been so heavily altered by the effect particularly of wave action and longshore sediment transport, and the river valleys have been drowned by rising sea level.

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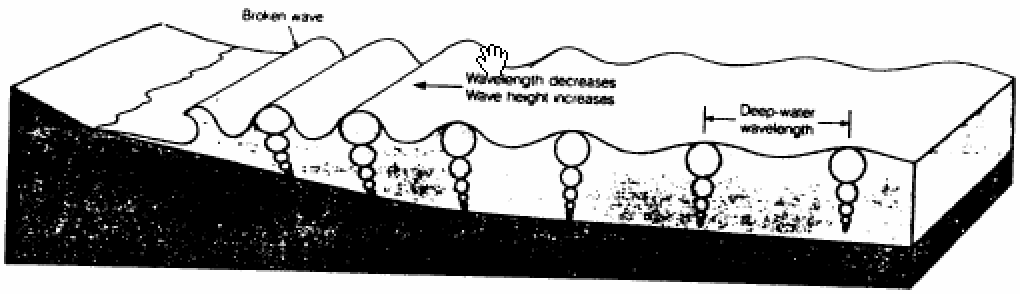
The remaining figures shown below illustrate the processes associated with straightening of shorelines in response to wave erosion and coastal sediment transport, as described on previous pages.



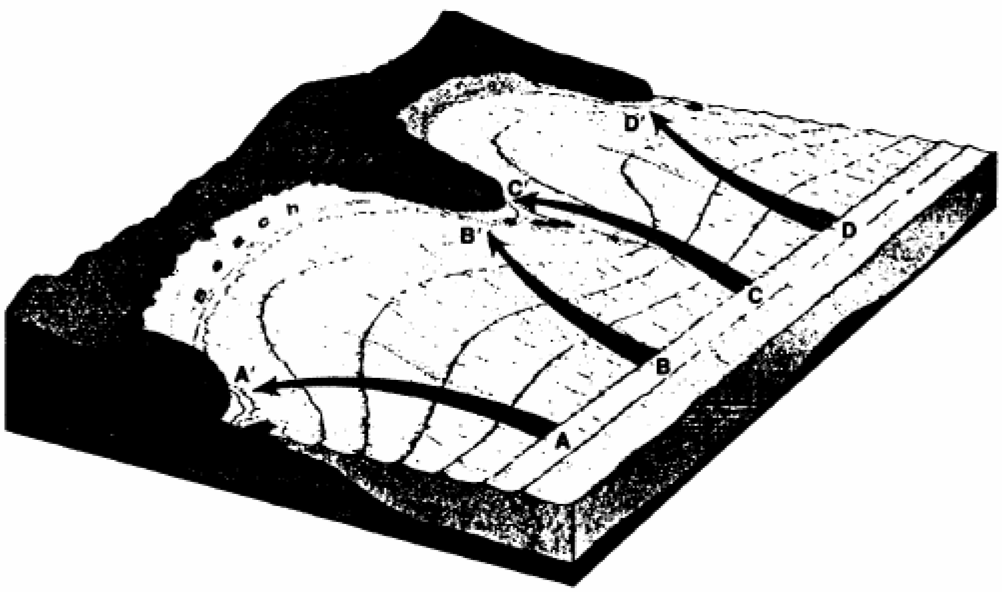
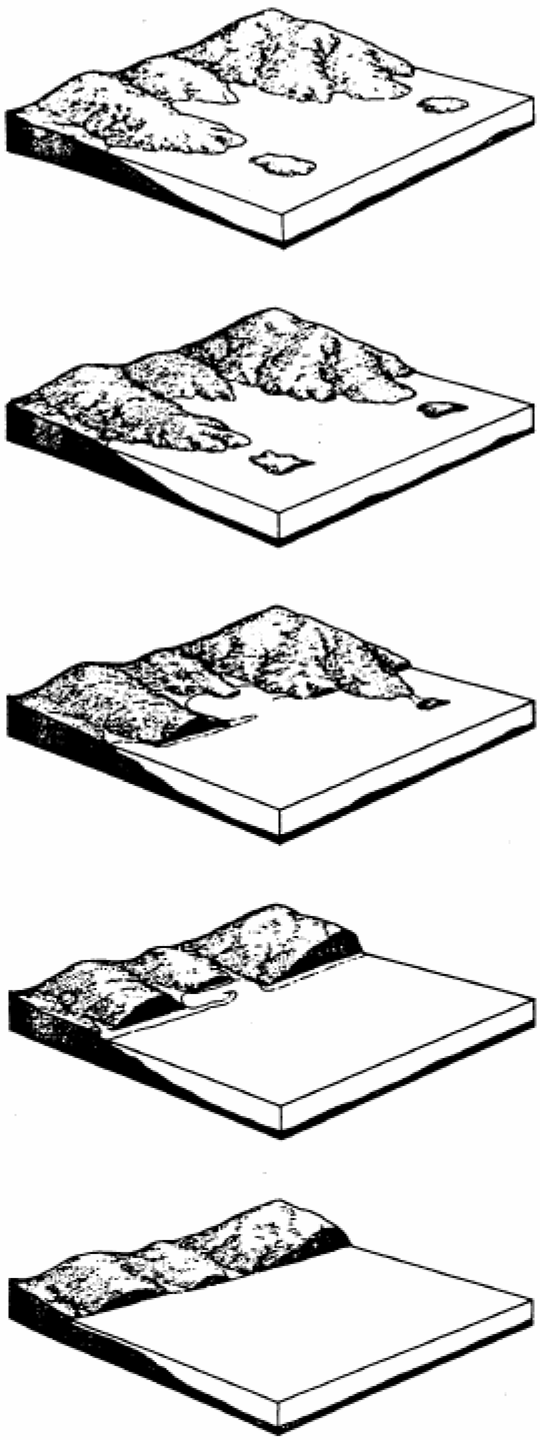


FIGURE 16.5
Wave refraction concentrates energy on headlands and disperses it across bays. Each segment of the unrefracted wave, AB, BC, and CD, has the same amount of energy. As the wave approaches shore, segment BC encounters the sea floor sooner than AB or CD and moves more slowly. This difference in the velocities of the three segments causes the wave to bend, so that the energy contained in segment BC is concentrated on the headland, while the energy contained in AB and CD is dispersed along the beach.

On the page after next is a map from ma 2000 publication illustrating the distribution of lands vulnerable to sea-level rise.



(A) A rise in sea level floods a landscape eroded by a river system and forms bays, headlands, and islands.

(B) Wave erosion cuts cliffs on the islands and peninsulas.

(C) Wave-cut cliffs recede and grow higher, and headlands are eroded back to a sea cliff. Sediment begins to accumulate, forming beaches and spits.

(D) Islands are completely eroded away, beaches and spits enlarge, and lagoons form in the bays.

(E) A straight shoreline is produced by additional retreat of the cliffs and by sedimentation in bays and lagoons. The large wave-cut platform then limits further erosion by wave action.

FIGURE 16.21
The evolution of a shoreline of equilibrium from an embayed coastline involves changes due to both erosion and deposition. Eventually, a smooth coastline is produced, and the forces acting on it are essentially at equilibrium, so neither erosion nor deposition occurs on a large scale.

385

Present-day rate of sea-level rise: http://www.agu.org/pubs/crossref/2004/2003RG000139.shtml