Influence of the Geologic Framework on Spatial Variability in Long-Term Shoreline Change, Cape Henlopen to Rehoboth Beach, Delaware

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ABSTRACT


Delaware’s coastal zone, both above and below the water line, is a mosaic of relict Delaware Bay and Atlantic shorelines created during different highstands of global sea level. Sediments deposited during oxygen-isotope Stage 5 (approximately 125,000 to 80,000 years before present) and in earlier-Holocene transgressive coastal environments (e.g., lagoon/estuary, estuarine beach, tidal inlet) are currently eroding in the shoreface as the modern shoreline migrates landward and upward through time. In coastal regions that lack significant input of new sediment to the littoral system, the local geologic framework can provide an important context for the interpretation of spatial and temporal trends in long-term erosion. Even where geologic data are abundant, the difficulty lies in providing information synthesized in a manner readily usable by coastal zone managers.

High-resolution seismic-reflection profiles, cores, and historical shoreline positions were used to refine the geologic framework and explore ways in which the framework influences the geomorphology and long-term retreat of the modern beach system along the Atlantic coast of Delaware. Examples are presented from Cape Henlopen, where a Holocene sand unit is supplying enough beach-compatible sediment to slow shoreline retreat, and Rehoboth Beach, where a Pleistocene headland appears to be more resistant to erosion than similarly composed Holocene sediments to the north. Visual correlation of long-term shoreline-change rates with a generalized geologic framework indicates that, statewide, the lowest relative erosion rates occur where relict Pleistocene or Holocene shorelines intersect the modern beach system. Geostatistical analyses indicate that although it may be possible to quantify the effects of large-scale transitions in the geologic framework, smaller-scale phenomena are not resolvable with this method. The results of this study can prove useful to coastal managers, scientists, and engineers by providing a scientific basis for modifying methods to calculate past and predict future shoreline change, and a mechanism for incorporating geologic data into the evaluation, design, and monitoring of beach nourishment and other coastal management activities.

ADDITIONAL INDEX WORDS: Antecedent geology, beach erosion, seismic stratigraphy, geostatistics, hazard identification.

INTRODUCTION

Given the tremendous economic investment in the coastal zone and the severity of the natural hazards that threaten this investment, models that predict future shoreline position are essential to effective coastal zone management and hazard mitigation. Ideally, such models would incorporate changes occurring on annual and decadal scales, the time scales at which coastal zone managers and land-use planners must operate. Some numerical models developed for engineering applications have been used to predict change over intervals of hours to years, from the erosion of an area caused by an individual storm to the redistribution of beach nourishment fill across the shoreface (Kriebel and Dean, 1985; Hanson and Kraus, 1989; Larson and Kraus, 1989; see discussions in Naval Postgraduate School, 2000; Thieler et al., 2000). Similarly, geologic models have been used to explain changes occurring over decades to millennia (Kraft, 1971; Fletcher et al., 1990). These process models and paleogeo-
graphic reconstructions tend to emphasize regional processes, not changes likely to occur at a specific site or within a specific period of time. Each engineering and geologic model approach has strengths and weaknesses owing to the original question it was designed to answer. A comprehensive model aimed at providing predictions suitable for coastal management and hazard mitigation must incorporate the appropriate aspects of the individual engineering and geologic models, and consider all of the site-specific factors that influence coastal evolution.

The driving forces behind shoreline change operate over a variety of temporal and spatial scales. Among the long-term, natural causes, sea-level rise and sediment supply are usually the most significant. The vertical and spatial range over which sediment-transport agents like wind, waves, and currents act is fundamentally determined by relative sea level. The retreat of continental glaciers during the Holocene Epoch has led to a eustatic rise in sea level of 120–125 m since 20 thousand years ago (kya) (Fairbanks, 1989), although melting of the ice sheets was complete by approximately 4–5 kya (Douglas, 2001). Factors such as glacial isostatic adjustment (GIA), thermal expansion of the surface ocean, melting of polar ice caps and alpine glaciers, and storage/mining of surface waters also affect global sea-level trends (Douglas, 2001). Sediment supply can magnify or counteract shoreline-change trends related to fluctuating sea level. For example, even in a period of rising seas, the shoreline on a terminal spit may move seaward due to a prolonged influx of sediment delivered by longshore transport. Conversely, shoreline retreat may be magnified where sea level is rising but little or no sediment is delivered via longshore transport. The two primary sources of sediment to the nearshore system are upland rivers and erosion/rewiring of existing sediments. Along the U.S. mid-Atlantic coast, most river-borne sediment is trapped by dams, or by estuaries and fringing marshes, which results in a sediment-starved shoreface and shelf.

A variety of short-term processes punctuate the record of long-term shoreline change. Dramatic, short-term changes are prevalent in areas near tidal inlets due to cycles of opening, closing, and migration of inlets over time scales of years to decades (Fitzgerald et al., 1978; Hayes, 1979; Oertel, 1985; Galgano, 1998). Nearly 65% of the Atlantic shoreline from New York to South Carolina is influenced by tidal-inlet processes (Galgano, 1998). Shoreline positions also vary seasonally due to differences in summer and winter wave climates (Smith and Zarillo, 1990; Pajak and Leatherman, 2002). Storms are the most dramatic agents of shoreline change, both in terms of absolute shoreline position and displacement of sediment. They are short-duration, low-frequency, high-energy events that exert considerable control over the evolution of coastal landforms (Moody, 1964; Morton et al., 1995; Douglas et al., 1998). Individual storms or periods of intense storm activity can cause tens of meters of shoreline recession over several days, equivalent to 50 or 60 years of gradual retreat related to sea-level rise (Douglas et al., 1998; Zhang et al., 2002). Periods of increased storm activity corresponding with large-scale climatic and oceanographic processes such as the El Niño/Southern Oscillation (ENSO) result in dramatic changes to the coast. ENSO events are known to produce sea-level anomalies that raise water levels tens of centimeters for 6 months to a year (Komar and Enfield, 1987), which allow damaging waves to attack higher along the shore. Increased storm wave power and wave height during ENSO events also contribute to significant beach erosion (Komar and Good, 1989; Griggs and Brown, 1998; Kaminsky et al., 1998).

Given the variety and complexity of the governing physical processes, developing models to explain past and predict future shoreline change is challenging. Historical shoreline datasets may extend back 150 years or more, but recorded shoreline positions are not ideally spaced through time; one or two positions may be available from the 19th century, two or three more represent the early- to mid-20th century, while four or more positions may have been gathered over the last 20 years. Each position represents a snapshot of the shoreline driven by conditions immediately preceding the measurement; thus, each reflects different components of the various change processes. For example, shoreline positions are sometimes measured immediately following major storms to document damages. Although the shoreline may recover to a position near its pre-storm condition (Morton et al., 1994), the historical dataset rarely resolves the recovery process. Overall, care must be taken during estimation of long-term shoreline-change rates to examine individual data points and, when justifiable, reject those that are not representative of long-term processes. A number of studies have examined these issues (Tanner, 1978; Smith and Zarillo, 1990; Galgano and
Leatherman, 1991; Crowell et al., 1993; Byrnes and Hiland, 1995), as well as the use of end point, linear regression, or other methods to calculate rates (Dolan et al., 1991; Fenster et al., 1993; Crowell et al., 1997; Galgano et al., 1998; Honeycutt et al., 2001) and use of post-storm shorelines (Fenster and Dolan, 1994, 1999; Douglas et al., 1998; Douglas and Crowell, 2000; Honeycutt et al., 2001; Fenster et al., 2001; Douglas et al., 2002).

In this paper, we present both qualitative and quantitative approaches for explaining historical shoreline change that rely principally upon geologic information and geostatistics (i.e., the theory of regionalized variables), respectively. Two case studies from the Atlantic coast of Delaware are provided to quantify the degree to which the nearshore geologic framework has influenced spatial (alongshore) patterns of shoreline change. The implications of this relationship on predictions of future shoreline change and identification of erosion-hazard areas are also examined.

Geologic Setting

Erosional and depositional processes associated with numerous cycles of glacio-eustatic sea-level change have controlled the large-scale geologic evolution of the coastal zone in Delaware over the last 500,000 years (Figure 1). The modern transgressive shoreline along the Delaware coast is microtidal and wave dominated (Oertel and Kraft, 1994), characterized by a continuous line of barrier or headland beaches broken only by the stabilized Indian River Inlet (Figure 1). The innermost shelf and lower shoreface are characterized by the highest concentration of shoreface sand ridges observed on the U.S. east coast (McBride and Morrow, 1991). Extending southeast from Cape Henlopen, Hen and Chickens Shoal (Figure 1) is an ebb-tidal feature composed of well-sorted fine sand winnowed from cape-associated littoral drift and transported offshore by buoyant outflow from Delaware Bay (Kraft, 1971). Although regional sediment transport on the continental shelf is to the south, net littoral transport along the Delaware coast reverses to the north from a nodal point located a few miles north of the Maryland-Delaware state line (Belknap and Kraft, 1985).

The uplands within several kilometers of the shoreline are composed of Pleistocene and older coastal sediments, which include a series of shorelines trending to the northeast and oblique to the modern system (Kraft, 1971; Demarest et al., 1981; Demarest and Leatherman, 1985; Fletcher et al., 1990; Oertel and Kraft, 1994). Most of the sediment being eroded and transported in the littoral zone today is reworked from Holocene, Pleistocene, and Tertiary units, with minimal input of new, coarse material (Kraft, 1971; Demarest and Leatherman, 1985). The subsurface of the nearshore zone is characterized by a series of shore-perpendicular incised valleys that were downcut during the last glacial maximum and subsequently infilled with silts and muddy sands during transgression (Kraft, 1971; Belknap and Kraft, 1981; Fletcher et al., 1990). These valleys are separated by interfluvies composed of materials ranging from silt to coarse sand and pebbles (Kraft, 1971). Preservation potential of these units varies greatly, with incised-valley deposits preserved preferentially as shoreface erosion removes material down to a depth of 11–13 m; incised valleys often extend to 20–25 m (Belknap and Kraft, 1981, 1985). In the study area, the primary interfluve of interest is herein referred to as the Rehoboth Headland (Figure 2), which is taken to represent the upland comprising Pleistocene and older sediments (Kraft et al., 1979; Demarest et al., 1981) located north of Rehoboth Bay.

Cape Henlopen is a simple spit that marks the southeastern limit of Delaware Bay (Figure 2), with sediment supplied from the eroding Rehoboth Headland to the south. Over the last 2,000 years, Cape Henlopen has evolved from a recurved spit complex, to a cuspatate foreland with a beach accretion plain, to a simple spit following construction of two breakwaters near the mouth of Delaware Bay in the 1800s (Kraft, 1971; Kraft et al., 1978) (Figure 3A). Radiocarbon dating of materials from archaeological sites and sediment cores constrains the formation of the recurved spit complex to 2,000 to 500 years before present (ybp). Spit sands encircled a large area north and east of the Rehoboth Headland, transforming an estuarine environment into a sheltered lagoon; spit sands were subsequently deposited on top of the lagoonal sediments. At 500 ybp, the recurved spits merged with the Rehoboth Headland near the present-day location of the town of Lewes, forming the cuspatate foreland morphology until construction of the breakwaters (Kraft, 1971; Kraft et al., 1978).

Today, Cape Henlopen continues to prograde to the north and northwest at more than 5 m/yr, while the Atlantic shoreline between the cape and
the Rehoboth Headland retreats at an average rate of 2 m/yr (Galgano, 1998). Despite the rapid westward migration of the Atlantic shoreline, a prominent bulge in the shoreline is visible near the eastern terminus of the recurved spit complex (Figure 3B), in contrast to the indentation that exists adjacent to Gordons Pond. The Rehoboth Headland forms a more subtle bulge farther south along the shoreline.

Role of the Geologic Framework

In its review of the U.S. Geological Survey’s Coastal and Marine Geology Program, the National Research Council (1999) recommended that the framework geology of the coastal zone be mapped nationwide, and a predictive capability based on this knowledge should be developed to assist policymakers in dealing with coastal erosion.
Figure 2. Location map showing distribution of data, Cape Henlopen and Rehoboth Beach. Solid lines represent Chirp seismic-reflection survey tracklines. Dashed lines are tracklines of the seismic survey conducted by the U.S. Army Corps of Engineers. Filled circles mark core locations. Contour interval of bathymetric data is 10 feet.

and other natural hazards. What is it about the geologic framework that prompts such a high-priority status for national scientific research and related public policy decisions?

On a fundamental level, the antecedent, or inherited, geologic framework is the land surface that is being inundated and eroded during transgression. The resulting morphology and along-shore patterns of shoreline change are, to varying degrees, dependent upon the antecedent topography, distribution of sediment types across that surface, and rate of sea-level rise. Variations in sediment composition, density, cohesiveness, and resistance to erosion and transport will affect beach width, shoreface profile, and recession rates (Riggs et al., 1995). At their study sites in North Carolina, Riggs et al. (1995) found the highest shoreline-recession rates were associated with areas of old inlet and channel-fill structures, and along zones updrift and downdrift of obliquely oriented, subaqueous sand shoals. Areas characterized by hardbottoms or relict, resistant Pleistocene and Tertiary units, as well as sediment-rich Holocene shoal systems, exhibited slower rates of beach erosion.

Although underlying topography and sediments influence coastal morphology and shoreline-change rates, the work to erode and transport sediment is done by wind and water. Relative sea-level change, waves, and currents are essential elements for establishing links between the antecedent geologic framework and the migration of the shoreline. In evaluating the Texas Gulf coast, Morton (1979) attributed spatial variability in erosion rates to wave phenomena. The highest erosion rates occurred at headland apexes, where wave energy was focused and rapidly eroded the mud-dominated deltaic sediments; low rates of
erosion or net accretion occurred in bights where former littoral cells once converged (Morton, 1979). Contributing factors to the patterns observed by Morton (1979) were relative sea level change (i.e., eustatic rise, local subsidence, and short-term, secular variations) and sediment-supply issues (i.e., decreased sediment discharge from rivers, impacts of coastal structures, and proportion of sand eroded from mud-dominated substrates). Other studies have examined spatial correspondence between wave phenomena, nearshore bathymetry, and alongshore patterns of beach erosion (Moody, 1964; Hobbs et al., 1999), but unlike Morton (1979), these analyses did not consider the underlying geologic framework. Ideally, the best models of shoreline change will contain both geologic and oceanographic/hydrodynamic parameters, but meaningful interpretations concerning spatial or temporal patterns can still be derived from either set of parameters alone.

In this study, detailed nearshore geologic and geophysical data and a 150-year record of historical shoreline positions are evaluated to examine the influence of the antecedent geologic framework on spatial patterns in shoreline change along the Delaware coast. The functional hypothesis guiding this study is that antecedent geology principally influences shoreline-change rates in three ways: (1) variation in the erodability of sediments, particularly among units differing significantly in age or composition, will cause spatial discontinuities in retreat rates; (2) relict topographic highs (e.g., barrier beach, tidal delta) and lows (e.g., tidal inlets, fluvial channels) may retard or accelerate shoreline retreat, respectively; and (3) relict units similar in grain size/composition to the modern system may act as local sediment supplies, slowing shoreline retreat. As will be described in the case studies that follow, each of these factors can be demonstrated to have influenced spatial trends in shoreline change along the Delaware coast during historical times.

METHODS

To refine existing models of the nearshore geologic framework, high-resolution Chirp seismic-reflection profiles were collected along the Atlantic coast of Delaware and in the Lewes and Rehoboth Canal. The Chirp data were collected using an Edgetech X-Star Chirp Sonar System and model 216S towfish, deployed approximately 3 m below the sea surface, with a pulse frequency range of 2–10 kHz and a shot rate of 8 shots/second (i.e., a
trace spacing of approximately 20 cm). Horizontal control was provided through the boat’s DGPS navigation system. In addition, 3.5 kHz “ping” and limited 0.4–5.0 kHz “Boomer” seismic-reflection profiles were collected farther offshore by the U.S. Army Corps of Engineers (McGee, 1995). Seismic data were processed to remove the effects of ship motion (Honeycutt, 1997; Honeycutt et al., in review), but no other filtering or gain controls were applied. Vertical resolution of seismic-reflection data is largely dependent upon seismic source frequency. At a decimeter or less, the Chirp data provided the best vertical resolution of any of the available seismic-reflection profiles. Ground-penetrating radar (GPR) profiles collected near Cape Henlopen by Daly et al. (2002) were used to identify the subsurface expression of some depositional units. Existing core descriptions, archived by the Delaware Geological Survey or in various University of Delaware student theses and dissertations, provided ground-truth for definition of reflection events and seismic facies in the geophysical data. Individual source citations for cores used in this study are provided during presentation of the results.

To examine spatial patterns of long-term shoreline change, rates were determined from historical shoreline data compiled by Dr. Stephen Leatherman of the Laboratory for Coastal Research (LCR) at Florida International University and by Galgano (1989, 1998). The LCR database for the Atlantic coast of Delaware spans from 1845 to 1997, and comprises positions from National Ocean Service (NOS) T-sheets, aerial photographs, and recent GPS surveys. Data were incorporated into a digital medium and corrected for datum changes, distortion, or other errors (for a more detailed description of this process, see Crowell et al. [1991] and Galgano [1998]). According to Crowell et al. (1991), the maximum spatial error for NOS T-sheet shorelines ranges from 8.9 m (pre-1880) to 6.1 m (1930–1970), including survey, cartographic, digitizing-hardware, and human errors (Galgano, 1998). Error for shorelines derived from aerial photographs and GPS surveys are 7.5 m (Crowell et al., 1991) and 1.5 m (Morton et al., 1993), respectively. Shoreline-change rates were calculated from transects oriented perpendicular to the modern shoreline and spaced at various intervals, using simple linear regression applied to all non-storm-influenced shorelines (i.e., excluding data from 1929, 1962, 1970, and 1993). In light of the theoretical spatial errors described above and the temporal scale of the LCR database (> 150 yr), all shoreline-change rates derived from these data and reported here have an error of ± 0.1 m/yr (Galgano, 1998). Geostatistical methods applied to these shoreline-change rates will be discussed in conjunction with the results.

RESULTS

Geologic Framework of Cape Henlopen

Initial interest in the potential effects of geologic framework on spatial patterns of erosion was spawned from observations of the historical shoreline data adjacent to the recurved spit complex at Cape Henlopen (Figure 4). The shoreline in 1845 was relatively straight; if anything, only a slight indentation appears in the central portion of the shoreline shown in Figure 4. By 1882 and persisting through 1903, a prominent double-hump appears in the shoreline where there appears to have been little or no change in position, while erosion continued along areas to the north and south. This feature, although smoothed somewhat during the 20th century, persists through the most recent shoreline position measured in 1997 and is visible as the “bulge” noted in Figure 3B.

As discussed earlier, the evolution of the recurved spit complex has been documented in various studies (Kraft, 1971; Kraft et al., 1978), including a recent analysis by Daly et al. (2002) to examine the internal structure of the spits using GPR. A profile collected perpendicular to the axis of one spit (Figure 5) shows three prominent radar facies: (1) a basal reflection-free facies below –5 m, interpreted as estuarine mud that, because of the mud’s higher electrical conductivity and/or retention of saline pore waters, attenuates the electromagnetic signal; (2) a strong, subhorizontal reflection event at –5 m (base of the spit platform) and a more discontinuous event near –2 m that form the bounding surfaces for a facies characterized by sigmoidal-oblique reflections, which is interpreted to represent the subaqueously deposited spit platform sands; and lastly, (3) a facies characterized by concave-up, tangential-oblique reflections, which is thought to represent beach and dune sands (Daly et al., 2002). These bounding surfaces and facies were identifiable in GPR profile segments collected across an adjacent recurved spit, and thus can be considered representative diagnostic features.

Chirp seismic-reflection profiles were collected across the inner shelf and shoreface to map the
distribution of any preserved recurved spit sands (Figure 4). The profile shown in Figure 6 was collected in the vicinity of the 1882 Atlantic shoreline; although several meters of overlying material have been removed from the sea floor while this area has been exposed to shoreface erosion, sediment outcropping on the sea floor today is probably the same as or is similar to what was eroding in the shoreface in the late 1800s and early 1900s. Comparable to what was observed in Figure 5, a strong, fairly continuous, slightly northward-dipping reflection event (A) is traceable across most of the profile at a depth of 13–15 m below sea level; it underlies a thick sequence of sigmoidal-oblique reflections that extend up to the ringing of the seafloor reflection. This pattern is repeated in all of the shore-parallel profiles collected farther offshore (Figure 7), with the reflection-rich sediment packet located progressively farther to the southeast.

The seismic-reflection profiles in Figures 6 and 7 show a basal facies relatively free of reflections below event A. This acoustically transparent facies is only interrupted intermittently by reflection event B (Figure 7). KRAFT (1971) and KRAFT et al. (1978) interpret the facies beneath the cape spits as estuarine and/or lagoonal in origin. Cores JCK-DH-3-68, JCK-B1, and R4104 (JOHN, 1977) (Figure 4) provide lithologic data that substantiate the seismic-stratigraphic interpretations shown in Figures 6 and 7. South of these profiles, core R4104 (JOHN, 1977) shows a thick sequence (>10 m) of sandy clay and clayey medium sand with shell fragments, which records the change in fauna from estuarine or shallow-marine invertebrates (Ensis sp. and Anomia sp.) to lagoonal species (Mercenaria and Crassostrea) when the first cape spits encircled the area. The lagoonal unit is overlain by a surficial peat associated with the modern marsh. Near the south end of Figure 6, cores JCK-B1 and JCK-DH-3-68 (KRAFT, 1971; HALSEY, 1978) reveal a surficial medium sand overlying green clayey silt/silty clay with fine sand. The sand-silt transition in core JCK-DH-3-68 occurs at

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an elevation of approximately \(-14\) m, consistent with event A. Gas wipeout obscures a segment of event A in Figure 7, but such interference is not surprising given the likelihood of methane trapped in the underlying, organic-rich, lagoonal and estuarine sediments.

Based on interpretations of these and other shore-parallel profiles, as well as a shore-perpendicular profile that crosses them (Figure 8), a series of seismic units (I–IV) has been identified using similarities in the reflection character of the bounding surfaces and internal reflectors (or lack thereof). These units have been traced across the study area, with remaining profiles showing the spit complex extending beneath a basal horizontal reflector that underlies the modern Hen and Chickens Shoal. Seismic units I–IV are likely to be similar compositionally and are interpreted as representing separate pulses of sediment transport that formed the individual recurved spits; the distal portions of these spits are visible above the modern marsh surface on land (Figure 9). By the time units V and VI were deposited, lagoonal deposition was probably driven farther north and west. Because reflection events are weak in these units, subdivision of the sequence and correlation with specific recurved spits in the marsh were not possible.

The depositional units identified in the seismic-reflection profiles allow the calculation of a conservative estimate of the volume of sand that makes up the remaining subaqueous portion of the recurved spit complex. Using the thicknesses for units II–IV (cross-sectional area of almost 3,300 square meters) and an estimated north-south coverage of 1.5 km, the sand volume is approximately 4.9 million cubic meters. For comparison, the net annual longshore transport at Indian River Inlet...
Figure 6. Inner shore-parallel Chirp (2-10 kHz) seismic-reflection profile (upper panel) and interpreted cross-section (lower panel; only major reflection events shown). Vertical exaggeration is 13×. Depths are from sea level, but not adjusted for tidal height. Sea-floor and sea-floor multiple (SFM) events are not labeled, with the lower limit of ringing of the sea-floor reflection represented by the dotted lines and Roman numerals used to denote seismic units. Events, from 12–15 m, represent the spit platform, where spit sands (facies characterized by dense, relatively strong, predominantly sigmoidal-oblique reflections) were deposited on top of acoustically transparent estuarine or nearshore marine silts. Approximate crossing point of shore-normal seismic profile (Figure 9) is also marked.
Figure 7. Outer shore-parallel Chirp (2–10 kHz) seismic-reflection profile (upper) and interpreted cross-section (lower). Vertical exaggeration is 11×. See caption of Figure 6 for additional explanation of features. Event A, the split platform, is assumed to continue across the area obscured by gas wipeout.
Figure 8. Outer shore-parallel Chirp (2–10 kHz) seismic reflection profile (upper) and interpreted cross-section (lower). Vertical exaggeration is 11×. See caption of Figure 6 for additional explanation of features. Reflection events used to divide seismic facies in this and preceding profiles dip offshore and continue eastward for another 500 m, beneath the western flank of Hen and Chickens Shoal (area not shown).
Figure 11. Shore-parallel Chirp (2–10 kHz) seismic-reflection profile (upper) and interpreted cross-section (lower) from offshore Rehoboth Beach and Henlopen Acres. Vertical exaggeration is 15×. See caption of Figure 6 for additional explanation of features. Relationships of seismic units VII–XI are not related to units shown in Figures 6–9. Seismic units and reflection events C through F are described in the text.
Figure 9. Interpreted spatial relationships between seismic units (I–VI) and subaerially exposed remnants of recurved spits. Historical shorelines from Figure 4 are shown for reference. Dotted lines are Chirp tracklines, with white plusses noting the emergence of key bounding surfaces at the sea floor. Possible alternate southern limit of facies I is represented by the dashed black line. Subdivision of unit V and correlation with recurved spit segments was not possible due to the subsurface sediments becoming acoustically homogeneous once the area became dominated by sand.

is estimated at 84,100 m³/yr (110,000 yd³/yr) (U.S. Army Corps of Engineers, 1991), which is roughly one-sixtieth of the spit complex’s sand volume. Despite the abundance of sediment that exists here, long-term shoreline-change rates are still on the order of –1.5 to –2 m/yr, with areas north and south of the spit complex eroding at 2 to 3 m/yr.

Geologic Framework of Northern Rehoboth Beach

As documented in previous studies (Kraft, 1971; Belknap and Kraft, 1981, 1985; Fletcher et al., 1990), the area north of the resort community of Rehoboth Beach marks the transition between a headland composed of Pleistocene and older coastal sedimentary units (Rehoboth Headland), and a generally northward-thickening wedge of Holocene sediments deposited on top of an irregular pre-transgressive surface. Based on several deep cores collected at nearshore and inland locations, Kraft (1971) and Kraft et al. (1978) suggested that the pre-transgressive surface drops down to an elevation of –20 m within 2 km of the northern subaerial limit of the Rehoboth Headland. The pre-transgressive surface was incised deeply by the lowstand rivers associated with present-day Holland Glade and an unnamed stream (herein referred to as Tidewaters Creek, the main channel of which is now occupied by a portion of the Lewes and Rehoboth Canal) (Figure 10). During early stages of the Holocene transgression, this area became part of the ancestral Delaware River estuary, and remained as such un-
til the high-energy Atlantic coastal environments migrated to that location (Fletcher et al., 1990).

Historical shoreline data from the northern flank of the Rehoboth Headland (Figure 10) show an interesting spatial pattern of differential rates of long-term retreat. While erosion appeared to be fairly uniform between 1845 and 1882, subsequent shoreline positions converged at the northern edge of the headland, adjacent to the residential community of Henlopen Acres. Several small groins were installed in the 1920s and 1930s to trap northward longshore transport in the area (Galiano, 1989). An increase in erosion north of the structures following their construction would have been expected due to interruption of the natural sediment supply. As a result, interpretations of the spatial pattern of shoreline change over the 150-year period of record must recognize the influence of these groins as an inseparable component of that change. Although there are no shoreline positions available after 1882 and before the groins were constructed, shore-protection structures are typically only built in response to an existing erosion problem. The question remains: why does there appear to have been greater erosion historically at Henlopen Acres and points north than at Rehoboth Beach?

Seismic-reflection data collected immediately offshore provide an explanation for this shoreline-change pattern (Figure 11). A series of distinctive seismic facies that are traceable laterally across the study area are interpreted from this and additional Chirp profiles. (Note: Although higher Roman numerals were used to label the seismic units in Figure 11, the ages of these units are unrelated to those established for the seismic units at Cape Henlopen.) A sequence of weak to moderately strong, horizontal reflections with intervening two- to three-meter thick, reflection-free sequences comprise seismic facies VII, which is interpreted
as the core of the Rehoboth Headland; this facies continues south for a distance of nearly 10 km. Core descriptions from KRAFT (1971) indicate that these Pleistocene units are generally made up of silt with fine-sand laminae and occasional layers of coarser sand. Little or no overlying Holocene material appears above the Pleistocene units in cores collected offshore, an observation corroborated by shore-perpendicular chirp profiles (not shown). A tangential-oblique, concave-up reflection (event B) truncates unit VII; this erosional surface is overlain by a series of weak, tangential-oblique reflections that continue upward until they are truncated by the sea-floor reflection and ring (unit VIII). As the reflections in unit VIII are weaker than or comparable in strength to those of unit VII, this unit has been preliminarily interpreted as Pleistocene in origin, possibly a remnant of a shoreface associated with oxygen-isotope Stage 5. Unit VIII (and possibly event B) is truncated to the north by another irregular but stronger reflection (event C) that is traceable down to a depth of 15 m, which is interpreted as the Holocene-Pleistocene sequence boundary.

The remaining reflection events and seismic units shown in Figure 11 are still being evaluated and correlated with existing core information. Preliminary interpretations indicate a complex Holocene evolution of the Tidewater Creek incised valley, characterized by early fluvial and/or estuarine deposition (unit IX) and multiple occupations of the valley by a tidal inlet. Incision is inferred from truncation events D and F, with infill represented by units X and XI, including the undulating event E, which is interpreted as a preserved sand wave. Cores JCK-3DH-70 (HALSEY, 1978) and R4108 (JOHN, 1977) collected inland of the modern shoreline (Figure 10) show lower valley fill to be composed of fine-grained materials (silt and a “stiff clay,” the latter described in R4108) that are consistent with estuarine depositional processes. These sediments are overlain by several meters of white to gray, medium to coarse sand with shell fragments and pea gravel, interpreted as inlet-related sediments.

Spatial Distribution of Shoreline Change

In the above examples, spatial variability in historical shoreline recession was shown over narrow, 1–2 km segments of the coast. These sites, and the spatial variations shown there, remain significant when placed into context with shoreline-change rates for Delaware’s entire 40-km shoreline. When calculating shoreline-change rates, what transect spacing is ideal for resolving significant spatial patterns of erosion and/or accretion? To explore this issue, long-term, linear-regression rates were calculated at various intervals along the Delaware coast, again avoiding the area near the rapidly prograding simple spit at Cape Henlopen. In addition, the data at Indian River Inlet and along its arc of erosion (GALGANO, 1998) are dominated by the short-term erosion caused when the inlet was stabilized. Despite the installation of a sand-by-passing system to restore much of the natural longshore-transport volume (U.S. ARMY CORPS OF ENGINEERS, 1991), evaluation of long-term trends for this analysis requires that the inlet-related rates be omitted from consideration. As shown in Figure 12, spacing transects every 5 km produces only eight rates, two of which are inlet-related and thus discounted. Erosion appears ubiquitous, but significant jumps in rates occur that would yield considerable uncertainty for hazard-area identification in intervening areas. At 2-km transect spacing, areas experiencing moderate erosion (< 1 m/yr) broaden, while erosion appears to worsen closer to the cape. At 0.5-km transect spacing, areas experiencing only moderate erosion broaden further; however, little or no discernable change in the erosion trends exists once the transect spacing is decreased to 0.01 km, suggesting that the point of diminishing returns with respect to effort expended has been reached for this segment of the coast and this shoreline dataset.

Given the local scale (1–2 km) of the case study examples, do these proposed mechanisms by which the geologic framework influences spatial patterns of long-term shoreline change hold when evaluating the coastal zone of the entire state? Shoreline-change rates calculated at 100-m intervals were plotted alongside a geologic cross-section by RAMSEY (1999), which was generated from cores collected on or landward of the modern beach/dune system (Figure 13). The lowest shoreline-change rates correlate visually with the surface distribution of Pleistocene units, namely at the Rehoboth and Bethany headlands. Moderate to high erosion rates correlate visually with Holocene sediments, with a slightly decreased erosion rate observable adjacent to the recurved spit complex.

Geostatistics

From a hazard-identification perspective, the above interpretations, which rely on potentially
subjective visual correlation, are of limited use in designating hazard-prone areas or in predicting future shoreline change. To be useful as a predictive tool in evaluating shoreline data and delineating hazard-prone areas, the influence of the geologic framework on rates of shoreline change, if any, must be established more quantitatively. Linear multiple regression was applied to determine the physical factors that control shoreline-change on Lake Ontario (AMIN and DAVIDSON-ARNOTT, 1997), however the geotechnical properties of the sediments (glacial till) were uniform across the study area and thus, the geologic framework was not an independent variable. DOLAN et al. (1992) used geostatistics to examine spatial patterns in shoreline-change rates on the North Carolina coast, but they only evaluated trends within the shoreline dataset; they did not evaluate geologic or oceanographic factors that may have influenced the results. Although their intent was not to determine causal relationships, the statistical techniques presented in DOLAN et al. (1992) are suitable for deciphering any spatial dependence within the shoreline dataset.

Following the methods described by DOLAN et al. (1992), shoreline-change rates for the Atlantic coast of Delaware (north of Indian River Inlet, but south of the prograding spit tip) were analyzed to discern the degree of spatial dependence. The spatial dependence of long-term shoreline-change rates on neighboring values was calculated using the semivariance function, which is defined as the average squared difference of values obtained at two locations (ISAACS and SRIVASTAVA, 1989). Semivariance values can be plotted as a function of the distance between values (the lag) in a graph called a semivariogram (Figure 14). As shown in this ideal semivariogram, the semivariance is, on average, near zero for all values measured a relatively small distance apart. However, the semivariance is expected to increase rapidly as the distance or lag increases, until spatial dependence no longer exists, i.e., the sill is reached. The lag at

Figure 12. Long-term shoreline-change rates along the Delaware coast, calculated using linear regression and at various intervals of transect spacing beginning about 2 km south of the prograding simple spit at Cape Henlopen. Gray points and corresponding line segments are rates measured within the 10-km Indian River Inlet arc of erosion, as determined by GALGANO (1988). Decreased transect spacing provides increased detail in the spatial distribution of shoreline-change rates. However, a point of diminishing returns for this dataset begins around 150 m, as spurious rates of accretion appear where no mid-19th century shoreline exists (27 km) and where an inconsistency in the surveying of the shoreline near the south jetty of the inlet (21 km) implies that there has been no net erosion over this timeframe.
Figure 13. Long-term, linear-regression shoreline-change rates for the entire Delaware Atlantic coast (upper) and the subsurface distribution of Holocene and older units along a line of section located just landward of the modern shoreline (simplified from Ramsey, 1999). With the exception of the accretion of the spit at Cape Henlopen, the lowest erosion rates correlate with areas where Pleistocene sediments exist at the surface, such as along the Rehoboth and Bethany headlands. Aside from the Indian River Inlet and its arc of erosion, the highest erosion rates occur along the Holocene deposits located south of the cape, with a slight reduction in erosion adjacent to the recurved spit complex.

which the sill value is obtained is called the range of the dataset.

Using shoreline-change rates calculated from transects spaced at 50-m intervals, the semivariogram depicting spatial dependence for the northern portion of the Delaware coast is shown in Figure 15. Commonly, stationarity (a mean of zero and variance of one) is induced in a dataset using any of a number of mathematical transformations prior to spatial analysis. However, the slopes of the semivariograms for raw (Figure 15, upper graph) and log-transformed (Figure 15, lower graph) rates in this study were essentially identical, as were the results for other transformations. Unlike the ideal semivariogram, rates remain strongly dependent for a considerable distance between values, until the semivariance increases rapidly at a lag of approximately 3 km and finally peaks (or reaches a sill) at a lag of 8 km. The apparent decrease in semivariance between 9 and 11.5 km is likely an artifact associated with the analysis being carried to a sample size of one.

**DISCUSSION**

As identified through the geophysical data, the relict, recurved spit complex south of Cape Hen-
lopen occupies a 1- to 2-km wide segment of the Atlantic coast, with sediment eroding in the shoreface that is roughly equivalent in composition to what is being transported along and across the littoral zone today. To the north and south of the complex, shoreface erosion is cutting into a variety of materials, ranging from fine-grained lagoonal sediments near Gordons Pond, to coarser materials deposited in earlier-Holocene estuarine beach and tidal inlet settings, and later-Holocene cape-related environments. In terms of the mechanisms by which the geologic framework can influence shoreline-change rates, the recurved spit example would correspond to factor 3, localized sediment supply. The nearly 5 million cubic meters of sediment that remain buried in the nearshore zone, in combination with an as-yet unquantified volume of sand stored in the subaerial spits, has slowed (but not halted) long-term erosion. To a lesser degree, the topographic relief between the subaerial/subaqueous spits and the Gordons Pond relict embayment could also contribute to the differential rates of erosion (factor 2). The 10-foot bathymetric contour deflects seaward slightly in the vicinity of seismic units I–V, but the seismic profiles show no vertical offset in the sea floor between the spit sands and lagoonal deposits to the south.

The Rehoboth Headland spans approximately 7.5–8 km of the Atlantic shoreline, the northern 4 km as a headland beach, the southern 3.5–4 km a barrier beach fronting Rehoboth Bay. Regardless of beach morphology, shoreline-change rates fronting the Rehoboth Headland are 0.5 to 1.0 m/yr, with a rapid increase to 2 m/yr north of the headland. The composition of the Pleistocene and Holocene materials—fine-grained silts with sand laminae and silt/clay overlain by medium to coarse sands, respectively—should, all other sediment properties being equal, produce the opposite relative rates of shoreline change; the sand would act as a local sediment supply, while the fine-grained materials would be winnowed away in the high-energy shoreface. However, the other sediment properties are not equal, as the Pleistocene deposits are considerably older and were subject to compaction and/or particle cementation by groundwater during subaerial exposure. In this case, differential susceptibility to erosion (factor 1) is the mechanism controlling shoreline-change rates. Topographic relief between the Rehoboth Headland and the Tidewaters Creek/Holland Glade incised valley system to the north may also play a role in the rate variability (factor 2), but no clear offset of bathymetric contours or vertical changes in the sea floor on the seismic profiles are observed.

In Delaware, the semivariance analysis revealed that the spatial scale at which shoreline-change rates are dependent upon one another is equivalent to the spatial scale of the geologic units eroding in the shoreface. Spatial dependence was demonstrated to a lag of 8 km, which represents the difference between the distribution of Pleistocene and Holocene sediments across the study area. Although impacts of the localized sediment supply at the Cape Henlopen recurved spit complex are visible in the historical shoreline positions and shoreline-change rates derived from those positions, the semivariance function does not appear to be sensitive enough to resolve the feature. Thus, the relative impact of the localized Holocene sediment supply (with or without contribution from variations in antecedent topography) on shoreline change is much less than the impact of the variable resistance of Pleistocene versus Holocene units, even if the units have similar sediment composition.

Do these results mean that shoreline-change
rates need only be calculated every 8 km along the coast? As demonstrated in Figure 12, valuable information on the severity of erosion and the spatial limits of relatively stable stretches of coast are lost when the sampling interval exceeds a few hundred meters. Dolan et al. (1992) outlined procedures for identifying optimum transect spacing based purely on statistical analysis of the shoreline-change data, without geologic context. They found that a spacing of 265–625 m would provide shoreline-change rates within the 95% confidence interval of the actual rates at unsampled locations, while a spacing of 160–315 m provided 99% confidence in the measurements. Future geostatistical analyses of the Delaware coast (discussed below) will examine optimum transect spacing in light of the geologic framework and other pertinent parameters in those areas not dominated by inlet- and structure-related processes.

Ultimately, the goal of this combined geological and geostatistical research effort is to determine the spatial correlation of two variables simultaneously, namely the shoreline-change rate and the nearshore geology. Similarly, provided that sufficiently detailed data were available, the spatial correlation of shoreline-change rates or nearshore geology with nearshore oceanographic parameters would also be useful in understanding the process-response models at work. A cross-variogram can be used to compute the spatial co-dependence of two variables, making it the ideal tool for addressing the overall research goal. However, the difficulty in establishing this link between geology and shoreline-change rates (or any other continuous variable) lies in deconvolving a conceptual geologic framework into discrete variables for cross-variogram computations. Easily quantified sediment characteristics, such as mean grain size and shoreface slope, would be logical variables to examine; however, the lithology of relict shorelines, especially the recurved spits at Cape Henlopen, is similar to that of the modern system. Common sediment attributes will be used in developing the geostatistical model, but a number of other continuous and categorical variables are also being evaluated. Once several variables thought to accurately represent nearshore geology are determined, cross-variograms will be used to examine their spatial co-dependence with erosion rates.

Implications for Coastal Management

The interaction of myriad geomorphic, geologic, climatologic, and oceanographic variables along the coast results in numerous modes of shoreline behavior (Leatherman, 1993). Although long-term field experiments have been initiated to provide temporally robust datasets that link oceanographic measurements and shoreline positions (e.g., Pajak and Leatherman, 2002), the impacts of past water-level and wave-climate variations usually cannot be resolved in existing shoreline datasets. Nevertheless, scientists must evaluate even incomplete data to provide the hazard information that is essential to coastal management activities. The state of the science in erosion-hazard identification is to distill often complex shoreline-change analyses down to an annualized erosion rate with estimates of error and uncertainty. In applying these annualized rates to establish construction setbacks or for other management purposes, predictions of future shoreline position assume that changes will occur at the same rate or at the same magnitude in the same place. In areas where the geologic framework has been demonstrated to be a dominant influence on spatial and/or temporal patterns of shoreline-change, this assumption may or may not hold.

When the geologic units influencing shoreline change are oriented perpendicular to the modern shoreline (Figure 16A), areas of stability or instability will likely shift uniformly landward with the shoreline. In the case of Cape Henlopen, the width of the spit complex narrows in the north–south direction, which would trigger a slight northward extension of the zone of higher erosion (B versus B'). Similarly, the Rehoboth Headland is oriented roughly perpendicular to the modern shoreline, owing to the orientation of the Tidewaters Creek incised valley underlying Henlopen Acres. If the critical geologic units are oriented oblique to the modern shoreline, the stable zone (Figure 16B, zone E) will migrate as the shoreline moves landward through time. Simple extrapolation of shoreline-change rates perpendicular to the shoreline would, in this situation, erroneously identify the erosion-prone versus stable areas.

In addition to affecting predictions of future shoreline position, the geologic framework may be relevant when determining the appropriate methods and data to use to calculate rates of past shoreline change. As discussed previously, often a scientific basis exists for excluding shoreline positions that have been determined from a priori evidence to be unrepresentative of the long-term trend, such as avoiding intermingling of summer and winter shorelines or omitting positions mea-
Figure 16. Suitability of extrapolation of future erosion trends perpendicular to the modern shoreline in light of the orientation of the underlying geology. A: As seen in the Cape Henlopen case study, the recurved spit complex (Zone A) and Rehoboth Headland (Zone C) have historically provided relative stability (solid lines) when compared to the sand-limited area fronting Gordons Pond (Zone B, dashed line). In the near future, the locations of these zones of stability and instability will likely remain similar (Zones A', B', and C'), as the geologic units in question are oriented roughly perpendicular to the shoreline. B: In contrast, stability (Zone E) associated with a relict shoreline feature that is oriented oblique to the modern shoreline (D) will likely migrate south in the future.
sured after major storms (DOUGLAS et al., 1998; HONEYCUTT et al., 2001). In areas where the geology is a dominant feature influencing shoreline change rates, could shore-perpendicular variability in the eroding sediments render older shoreline positions unrepresentative?

As shown in Figure 10, the entire Rehoboth Headland eroded approximately 100 m between 1845 and 1882, but the shoreline has remained within 30 m of the 1882 position throughout the subsequent century. Shore-stabilization activities have been undertaken since the 1920s along the developed portion of the shoreline in Rehoboth Beach and Henlopen Acres. These activities consist of construction and maintenance of a groin field from 1922 to present, emergency post-storm beach renourishment in 1962, and additional beach renourishment during the 1990s in Rehoboth, as well as Dewey Beach to the south (unpublished data from the Delaware Department of Natural Resources and Environmental Control, provided by JEFFREY WAKEFIELD, Univ. of Delaware). If the 1845 shoreline is included in calculations of long-term shoreline-change rates, the quantity of annualized erosion is much greater than what has been observed over the last century and, more importantly, projections of future shoreline positions would overestimate the erosion likely to occur. Based on the Holocene paleogeography (FLETCHER et al., 1990; HONEYCUTT, 2003), a possible explanation for the observed decrease in erosion is that from 1845 until 1882, the material eroding in the shoreface could have been several meters of sediment dating from the early Holocene (e.g., freshwater wetland, fringing salt marsh, and/or estuarine beach environments) deposited above the Pleistocene units. Once shoreface erosion removed this Holocene material and began cutting into the compacted, oxidized Pleistocene units, the rate of horizontal shoreline retreat would have dropped to roughly what it is today.

Unfortunately, the critical a priori evidence necessary to justify omitting the 1845 shoreline from rate calculations near Rehoboth Beach, namely a map depicting surface distribution of Holocene sediments across the shoreface in the 1800s, is not available. Further, the impact of the shoreline stabilization activities cannot be separated from the annualized erosion rates. In the absence of these types of information, it may reasonable to use the remaining geologic evidence to justify down-weighting the 1845 position in the linear-regression analysis or in some other fashion. The critical point to be gathered from the examples provided in this study is that an evaluation of sediment variability perpendicular to the modern shoreline may ultimately prove to be as important to predicting future shoreline change as evaluations of variability in the alongshore direction.

CONCLUSIONS

Recent published research has shown that, generally, the most accurate models of past shoreline change are those that use linear regression and non-storm shorelines to calculate rates; these models are then considered the most reliable predictors of future shoreline position. The nearshore geologic framework is emerging as another factor to be considered when evaluating shoreline-change data. Visual correlation of shoreline-change and nearshore geologic datasets for Delaware shows that some of the lowest erosion rates are found in areas where relict shorelines intersect the modern beach system. These sediments appear to be more resistant to erosion or provide a steady supply of beach-compatible material. Semivariance analysis of shoreline-change data revealed spatial dependence of erosion data up to a lag of 8 km, which is equivalent to the spatial distribution of Pleistocene and Holocene sediments. A more robust geostatistical model is being developed to quantify the impact of the geologic framework on long-term erosion, but the greatest challenge lies in breaking down the framework into discrete variables. If this model can successfully explain residual spatial variability in shoreline-change rates (beyond the effects of stabilized inlets and other shore-protection activities), forecasts of future shoreline position can be improved in sediment-starved coastal areas. Where demonstrated to have a quantifiable influence, knowledge of the geologic framework can assist coastal managers and others involved in the identification and mapping of erosion hazards to mitigate against future erosion losses.

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