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Historical Shoreline Trends Along the Outer Banks, North Carolina: Processes and Responses

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ABSTRACT

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The shoreline rate-of-change statistic is calculated from sequential measurements of shoreline position. This statistic implicitly represents the cumulative impact of those processes which have influenced shoreline behavior. Knowledge of the phenomenological relationships between the oceanographic processes and the shoreline's response, however, is not required for the computation and utilization of rate-of-change statistics.

In this paper we suggest that an understanding of the processes governing shoreline behavior will greatly aid response-centered analyses. This will be true for numerous applications involving shoreline rate-of-change values, especially those which must determine the persistence of short-term deviations from the long-term shoreline trend. Unfortunately, process-response data from most of the world's coastlines are neither synoptic nor of high resolution. In addition, functional relationships between the processes and responses are difficult to quantify due to the synergistic nature of the shoreline processes.

For a 7.4 km reach along the Outer Banks, North Carolina, we demonstrate typical problems associated with identifying the principal causes of shoreline movement in a highly dynamic environment. When viewed over the time spans used in shoreline analyses, which utilize remotely-sensed data (≈10 to 150 years), the spatial continuity of the processes resulting in shoreline movement is limited to relatively narrow geographic segments along the shore. Thus, a single, long-term process, such as sea-level rise, does not appear to dominate shoreline movement over the 134 year record along the Outer Banks. Instead, relatively long-term trends in shoreline movement correspond to cyclic patterns in storm frequency and intensity, and short-term sea-level adjustments. Other processes affecting local sediment budgets, which can be difficult to quantify, include longshore variations in sediment transport and/or variations in the delivery and storage capacity of sources and sinks over time.

ADDITIONAL INDEX WORDS: *Shoreline trends, coastal processes, shoreline behavior, Outer Banks, North Carolina.*

INTRODUCTION

Shoreline movement is a complex phenomenon due to the synergistic nature of those processes which produce shoreline changes. Complications in deciphering spatial and temporal trends in shoreline movement can be partially attributed to the equivocal relationship between shoreline processes (causes) and shoreline response (effect) and partially to the difficulties involved in distinguishing long-term shoreline movement (signal) from short-term changes (noise). Even if the physics involved in each of the driving processes could be known completely, the impact of each individual process as well as the combined influence of many processes on shoreline dynamics would not necessarily be understood. Indeed, many of the numerical and analytical models used by coastal investigators to examine shoreline change are based on attempts to understand this complex process-response system (*e.g.*, HANSON and KRAUS, 1989).

In three companion papers, we have used and developed tools to investigate the spatial and temporal variability of shoreline movement (DOLAN *et al.*, 1991, 1992; FENSTER *et al.*, 1993). Using spatially-rich and relatively temporally-poor sample data from Hatteras Island, North Carolina, we created an optimal spatial sampling design and delineated historical, as well as future, temporal shoreline trends. In this paper, we attempt to link the spatial and temporal shoreline responses of a segment of the Outer Banks to their causative processes.

Although it is desirable to determine the relationships among processes and responses, such a determination is made difficult for a number of reasons. First, and perhaps foremost, shoreline studies suffer from a lack of consistent empirical data. Very few of the world's coastlines are systematically monitored. Wave statistics for North

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Carolina's 485 km long coast, for example, are computed from three offshore gauges (maintained by the National Oceanic and Atmospheric Administration; NOAA) and a few nearshore gauges. One exception to the coarse resolution of field data collection is the United States Army Corps of Engineers Field Research Facility at Duck, North Carolina, where a federal commitment enables the collection of simultaneous process-response data. Unfortunately, even this program is spatially (<2 km) and temporally (1980 to present) circumscribed. Moreover, process-response studies are limited by the quality and quantity of source data (*e.g.*, DOLAN *et al.*, 1991; MORTON, 1991), accurate representation of the shoreline (*e.g.*, CROWELL *et al.*, 1991), and by problems of analysis and interpretation (*e.g.*, MORTON, 1978, 1991; EVERTS and GIBSON, 1983; SMITH and ZARILLO, 1990; DOLAN *et al.*, 1991).

In this discussion we present an heuristic approach to understanding the processes which control shoreline response in the highly dynamic environment of the Outer Banks of North Carolina. This procedure illustrates the problems associated with understanding historical shoreline patterns and accurately predicting future trends.

STUDY AREA AND DATA BASE

For this study, we chose to analyze a relatively data-rich segment of the Outer Banks of North Carolina (Figure 1; DOLAN *et al.*, 1978). The response data were obtained from two historical National Ocean Service (NOS) T-sheets and 10 aerial photographs spanning the period from 1852 to 01 October 1986 (134 years). Shorelines were digitized at 50 m intervals within 3.7 km long, consecutively numbered and stacked, base maps.

From this data base, we randomly selected a 7.4 km long segment of coast (base maps 20 and 21) from the 62 km reach between Oregon Inlet and Cape Hatteras. The reach is located approximately 10 km north of Cape Hatteras and approximately 45 km to 52 km south of Oregon Inlet (Figure 1). This length of coast enabled us to investigate those problems typically associated with linking oceanographic processes with shoreline responses in a natural, dynamic, open-ocean setting and to compare and contrast local versus regional influences on shoreline dynamics. Furthermore, study of the processes within this coastal region coincides with and augments existing analytical research conducted in this area (*e.g.*, DOLAN *et al.*, 1991, 1992).

From base maps 20 and 21, we selected six transects for detailed, transect by transect, process-response analysis (Figure 1). DOLAN *et al.* (1992) showed that a minimum separation distance of 250 m (5 transects) to 600 m (12 transects) between any two transects can be used to estimate shoreline response (rate-of-change) at an unknown location to ± 1.0 m/yr with 95% confidence and to insure independence of adjacent measurements. In other words, information obtained from transects spaced at intervals shorter than these distances would be repetitive due to the high degree of spatial autocorrelation in shoreline response (verified by visual examination).

LONG-TERM TRENDS AND CAUSES OF TREND REVERSALS

Net shoreline change within a particular littoral cell is a function of either a sediment budget deficiency (or excess) for that region or the redistribution of sediment along a profile. The processes causing shoreline movement occur over a variety of temporal scales (Figure 2; MORTON, 1991; FENSTER *et al.*, 1993). These include sea-level rise (BRUUN, 1962), reduction or increase in sediment supply from inland sources (*e.g.*, JOHNSON, 1959), storm activity (*e.g.*, HAYDEN, 1975), longshore variations in wave energy (*e.g.*, MAY, 1983), seasonal profile changes (*e.g.*, WRIGHT *et al.*, 1979), and/or human interaction (*e.g.*, DOLAN, 1972). Clearly, determining the cause of a reversal in the long-term shoreline trend or of gross morphologic change will be easier in regions where fewer processes dominate shoreline response (*e.g.*, shifting ebb-channels and associated wave regime changes along inlet-dominated reaches; *e.g.*, FITZGERALD, 1976) or where human modification has occurred. The transects with the most difficult cause-effect relationships to discern are those located in regions where the processes producing the responses are synergistic and/or noisy.

For all but one transect, three distinct shoreline trends predominated along the coast fronting Avon, North Carolina (21-10) (Figure 3; FENSTER *et al.*, 1993). These trends were relatively consistent with those found for the Cape Henry to Cape Hatteras reach (averaged) by EVERTS *et al.* (1983) (Figure 4). Although EVERTS *et al.* (1983) showed spatial uniformity of temporal trends for shorelines north and south of Oregon Inlet, we have found that variations in temporal shoreline trends occur over relatively short distances in this region.

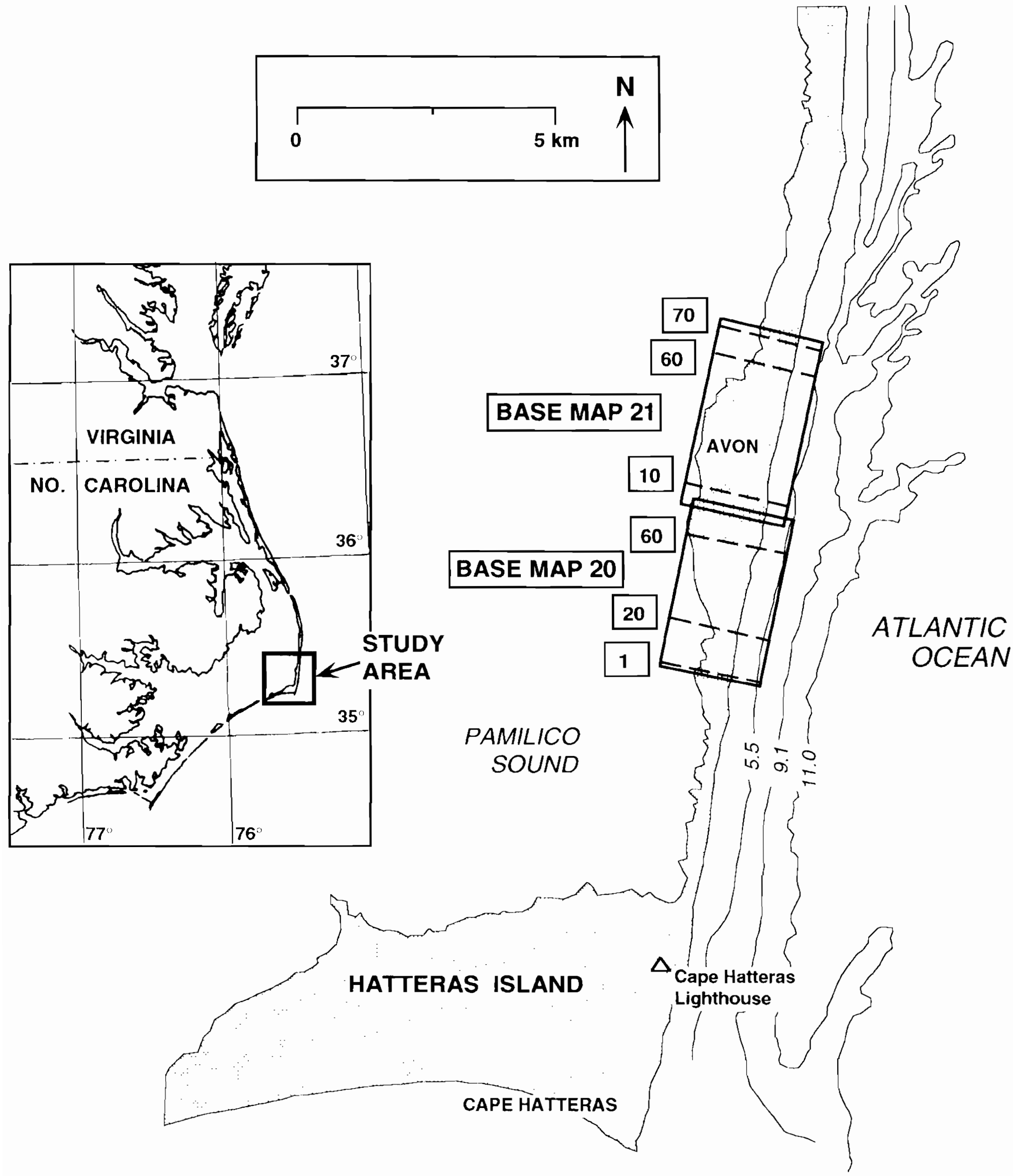

Figure 1. Map of study area: Outer Banks, North Carolina. Shoreline data are organized into base maps containing 72 transects spaced 50 m apart. Base maps (rectangles) and transects (dashed lines) used in this study are indicated on the map. Bathymetric contours are 5.5 m (18 ft), 9.1 m (30 ft), and 11 m (36 ft).

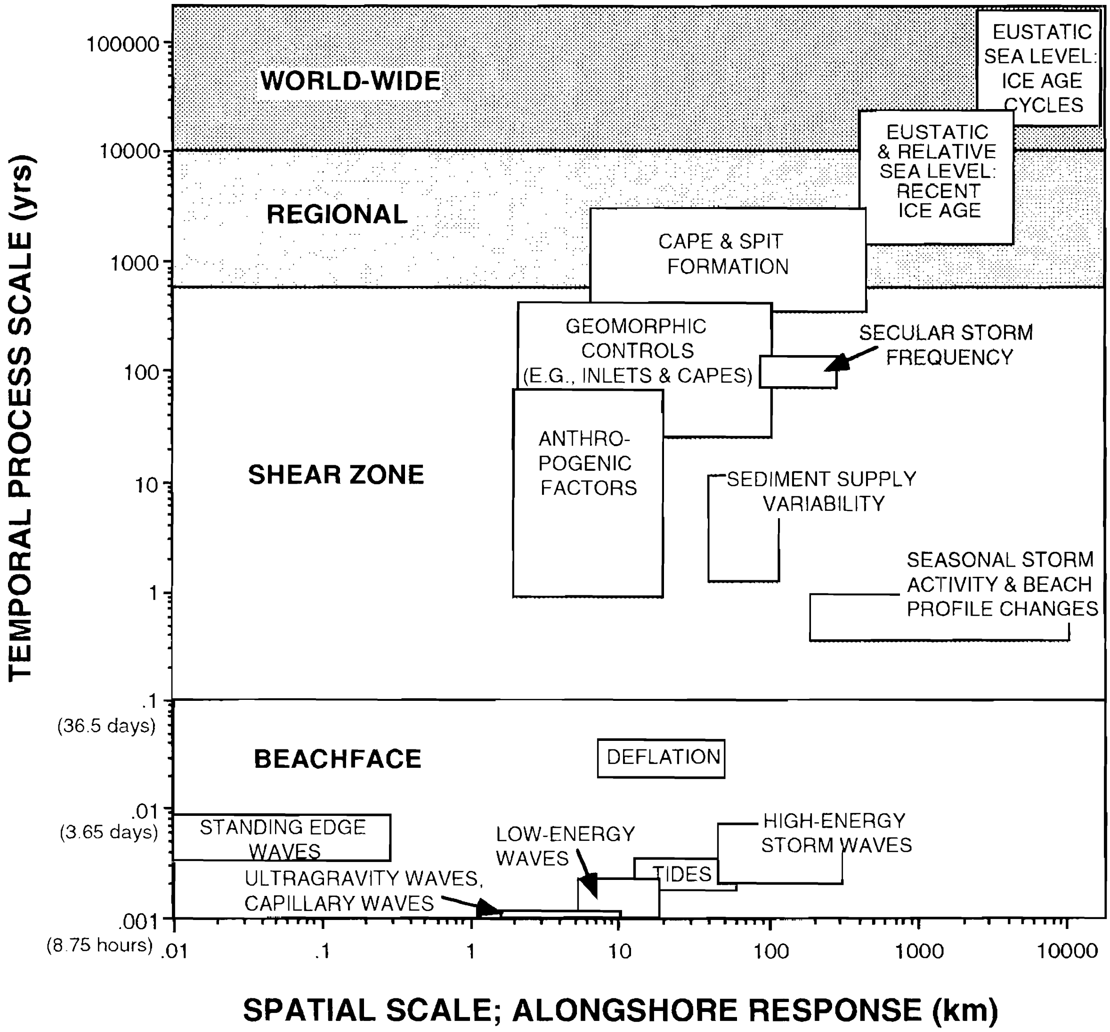

Figure 2. Temporal and spatial scales of coastal change. Each box relates the spatial (alongshore) impact of the various processes or anthropogenic factors responsible for shoreline movement to the time spans (longevity) of those processes.

These variations are an important consideration for comparing local beach dynamics to physical processes, since the processes occur over a range of spatial scales. For example, sea-level adjustments impact a much wider reach of coast than does wave energy, which diverges or converges due to local offshore bathymetric variations (Figure 2).

At Avon, North Carolina, shoreline movement at transect 21-10 is linear, indicating a constant and uniform long-term trend toward erosion at a rate of 0.4 m/yr (Figure 3). The transects located north and south of 21-10, however, display distinct third order polynomial trends (according to a Minimum Description Length criterion; FENSTER *et al.*, 1993). For the period 1852 to 1917, the transects south of 21-10 showed slight accretion (stability) with increasing erosion to the south ($>$50 m; $\approx$ $-$0.75 m/yr) while high accretion rates predominated north of 21-10 ($>$150 m; $\approx$ $+$2.3 m/yr) over this same time span. After 1917, however, all shorelines move in the same direction at a relatively similar rate. Erosion trends predominate until the late 1960's and early 1970's. At

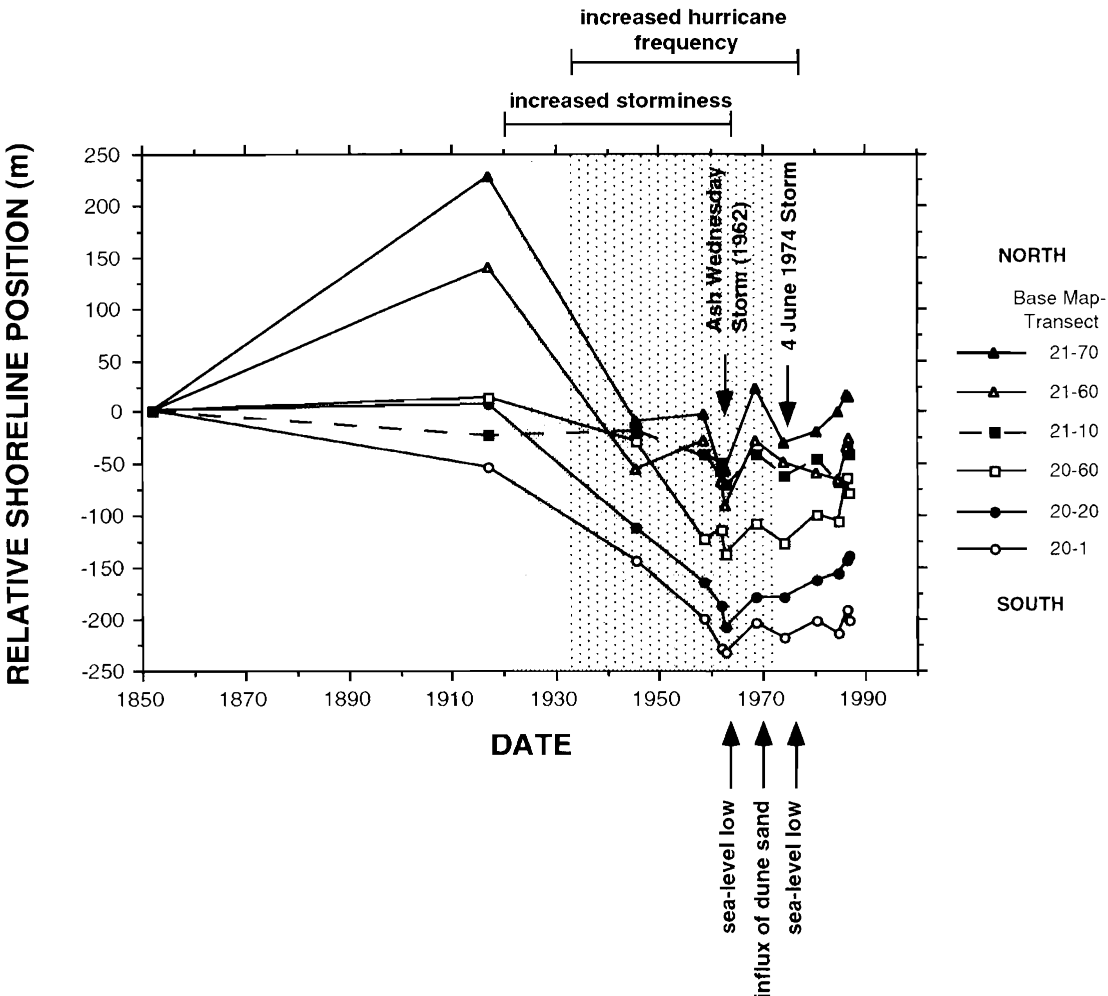

Figure 3. Shoreline position/time data for transects used in this study (locations shown in Figure 1) and the timing of coastal processes impacting shoreline movement along the Outer Banks. According to statistical analyses, all transects, except for 21-10 (dashed line), display nonlinear shoreline behavior. According to the trend analysis of FENSTER *et al.* (1993), the shoreline trend reversals occurred *c.* 1880–1920 and *c.* 1968–1972.

that time, a modern seaward shift in shoreline trend occurred and continued through 1986. In the following sections, we attempt to explain the causes of these trend reversals found along the 7.4 km Avon reach.

Oceanographic Regime: Sea Level

A three-step process is needed to determine the percentage of shoreline erosion that is related to sea-level rise. First, a reliable shoreline response model must be chosen; second, the model's terms need to be estimated; and third, the sensitivity of the terms must be tested.

Several two- and three-dimensional models are available for determining the relative impact of sea-level rise on shoreline movement (see review in KOMAR *et al.*, 1991). Most models are modifications of the BRUUN (1962) Rule, which assumes an upward and landward equilibrium profile translation in response to a rise in sea level, that the volume of sediment eroded from the shoreface is equal to the volume of sediment deposited on the ramp, that no sediment will be deposited seaward of closure depth, and that no alongshore transport will occur. Modifications to Bruun's two-dimensional (cross-shore) model include adjustments in order to depict more accurately the active zone of sediment transport (DUBOIS, 1977, 1990, 1992; DEAN, 1982; DEAN and MAURMEYER, 1983; KRIEBEL and DEAN, 1985; KRIEBEL, 1990), while three-dimensional models attempt to account for the longshore sediment gradient (EVERTS, 1985; HANDS, 1980, 1983; DEAN and MAURMEYER, 1983). The primary terms needed for testing the models are:

$$R = \frac{L_*}{B + h_*} S \qquad (1)$$

where shoreline retreat, R, is a function of the cross-shore distance, L_*, to closure depth, h_*, the berm height or elevation estimate of the eroded area, B, and the sea-level rise rate, S (BRUUN, 1962).

The second step in determining the percentage of shoreline erosion that is related to sea-level rise involves assessing the magnitude of S. The need for producing an accurate S term stems from the functional relationship of the Bruun Rule: Since $\frac{L_*}{B + h_*}$ reduces to $\frac{1}{\tan \theta}$, a relatively small increase in S results in a large shoreline retreat.

The tide gauge stations closest to Cape Hatteras are located at Wilmington, North Carolina; Portsmouth, Virginia; and Hampton Roads, Virginia. These gauges contain approximately 50 years of data ranging from 1936 to 1986 (LYLES *et al.*, 1988). The relative sea-level histories of the two regions north and south of Avon vary due to differences in vertical crustal movements (*e.g.*, GORNITZ and SEEBER, 1990; BRAATZ and AUBREY, 1987). Wilmington is situated on a structural high and is subject to continuing uplift along the Cape Fear Arch. Those gauges located in Chesapeake Bay are subject to high rates of land surface subsidence (possibly due to ground water withdrawal; DAVIS, 1987) or crustal subsidence (due to isostatic downwarping; BLOOM, 1967; SHEPARD and WANLESS, 1971). Thus, the Chesapeake Bay gauges record anomalously high rates of relative sea-level (RSL) rise whereas the Wilmington gauge records anomalously low rates of RSL rise (Figure 5). Regression analysis of the tide gauge records (uncorrected) gives RSL rates of 1.80 mm/yr at Wilmington, 4.30 mm/yr at Hampton Roads and 3.70 mm/yr at Portsmouth (LYLES *et al.*, 1988).

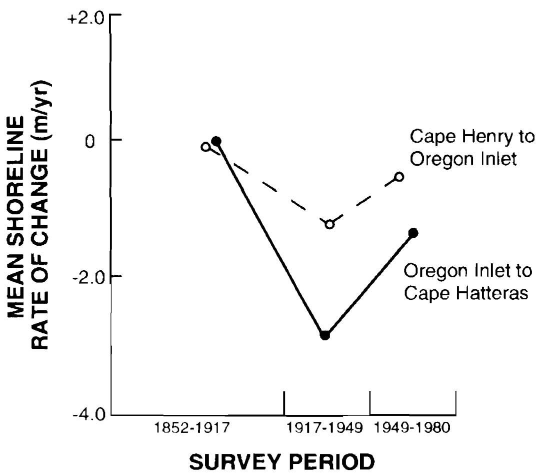

Figure 4. Temporally and spatially averaged shoreline rate-of-change values for the region from Cape Henry, Virginia to Cape Hatteras, North Carolina. Note the temporal rate changes from stability to erosion and finally towards a modern day lower erosion rate (After EVERTS *et al.*, 1983).

Plots of these data (normalized) show similar short-term trends in the relative magnitudes of average annual sea-level positions (Figure 5; HICKS and HICKMAN, 1988). Note, for example, the rise in sea level at all gauges from 1963 to 1972–1973. This rising trend was interrupted by one period of sea-level fall in 1968 and followed by sea-level fall from 1973 to 1976. Following 1976, the data become highly variable, alternating between relatively large magnitude rises and falls. The similarity among trends at stations located ≈400 km apart indicates that the physics governing sea-level movement at these stations is not localized. These processes could be related to shifts in the geostrophic mean westerly air flows between 35° and 45° north latitude, which result in vector changes of offshore winds; changes in sea-surface temperatures and subsequent thermal expansion; and/or unusually large contributions of fresh water. Shoreline movement occurring over relatively short time spans (the temporal domain of maps and aerial photographs), therefore, may be responsive to short-term variations in the long-term sea-level position. In any case, the S values, obtained by interpolation between the Wilmington and the Chesapeake Bay tide gauges, can be considered reliable. Although EVERTS (1985) used

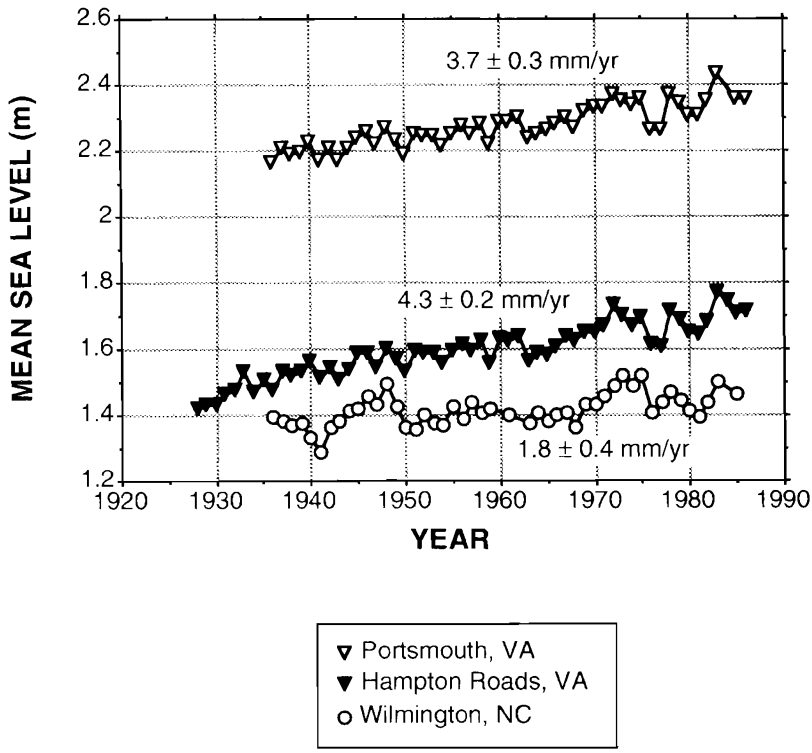

Figure 5. Uncorrected sea-level histories for tide gauge stations located nearest Cape Hatteras, North Carolina. Note the absolute magnitudes of sea-level trends vary widely, but short-term changes in the long-term trend are regionally distributed. Tide gauge data from LYLES *et al.* (1988).

values of +3.4–3.5 mm/yr for the Outer Banks, North Carolina, we used the value +3.8 mm/yr corresponding to the upper limit of S as interpolated by the NOAA's Tides Branch in collaboration with the Strategic Environmental Assessment Division (FRANK, 1992) and computed by DANIELS *et al.* (*in preparation*) for Nags Head, North Carolina (approximately 65 km north of Avon).

As expected, results for the various two-dimensional models are similar. Using 2 km for L_*, 20 m for h_*, and 2.4 m for B (tan $\theta \approx 0.012$), R is 0.34 m/yr (BRUUN, 1962; the Bruun Rule), 0.33 m/yr (DUBOIS, 1990; Transgressive Shoreface Model), and 0.35 m/yr (EDELMAN, 1972; considers diminishing dune height relative to the water level as the profile moves upward). We did not use the generalized Bruun Rule of DEAN and MAURMEYER (1983), which accounts for deposition on the barrier island or in the lagoon, since overwash is not a dominant process in the Avon area due to the presence of an artificial dune ridge system (constructed in the early 1930's; BIRKEMEIER *et al.*, 1984).

KOMAR *et al.* (1991) demonstrated the importance of expressing R in terms of a complete sediment budget by including a term (G_B) to account for the longshore gradient of littoral drift ($\partial Q_s/\partial Y$; where Q_s is the longshore sediment movement into or out of a control volume with alongshore length y and cross-shore width x), as well as contributions from aeolian, riverine, and offshore sources. When $\partial Q_s/\partial y$ is zero, the three-dimensional models reduce to two dimensions. The dissimilarity among the temporal trends over the Avon reach provides evidence that the assumptions $\partial Q_s/\partial y = 0$ and/or $\partial Q_s/\partial x$ = constant are not valid. EVERTS (1985) attempted to quantify the relationship between R and S using a sediment budget (within a control volume) approach for the Outer Banks, North Carolina. EVERTS (1985) accounted for the net quantity of sand reaching and leaving a particular reach by incorporating a term, V_0, to include all shoreline changes that are not due to sea-level rise:

$$V_0 = \Delta t \sum_{i=1}^{n} \frac{\Delta V_i}{\Delta t} \qquad (2)$$

where V_i is the volume of sand entering or exiting the control volume over time, and Δt is a result of n transport processes. EVERTS (1985) stated that the only important losses to the V_0 term for the Outer Banks are transport losses associated with Oregon Inlet (≈50 km north of Avon) and Cape Hatteras (≈15 km south of Avon). This conclusion implies that, over a relatively short reach between Cape Hatteras and Oregon Inlet (*e.g.*, 7.4 km along Avon), $\partial Q_s/\partial y = 0$. EVERTS (1985) does not, for example, account for cross-shore transport of sediment across the shoreface connected ridge systems (a component of $\partial Q_s/\partial x$) along Hatteras Island (Wimble Shoals and Kinakeet Shoals). Thus, acceptable explanations of shoreline trends resulting from sea-level variations are unfortunately restricted by an incomplete understanding of (and inability to quantify) spatially and temporally complex, three-dimensional sediment transport processes, such as those associated with inshore shoals.

It must be emphasized that sea-level rise models are most useful on an island-wide, regional scale and any attempt to quantify localized effects due to sea-level rise may not be accurate. Additionally, state-wide comparisons of observed shoreline rate of change values may not correspond to R as predicted by the Bruun Rule (DEAN, 1990; KOMAR *et al.*, 1991). For the North Carolina coast, however, the state-wide average (−0.6 m/yr) falls very near the Bruun "rule of thumb" proportionality of R = 100S. For North Carolina, therefore, longshore transport may be more uniform over the long-term than for other states. This may be due to the long linear spit and barrier island morphology positioned within a single littoral cell and to the few engineering structures located within the state. The Hatteras littoral cell extends ≈180 km from Cape Henry, Virginia, to Cape Hatteras, North Carolina, with volumetric losses associated with Oregon Inlet and Cape Hatteras (INMAN and DOLAN, 1989). Beaches located within this cell experience a net southerly sediment transport flux of 590,000 m^3/yr. Over short beach segments, such as the 7.4 km reach at Avon, variations in the temporal data indicate that the sediment budget may be influenced by local sediment sinks or sources (the G_B term, mentioned above, becomes more important) or by shorter-term temporal processes (*e.g.*, storm climate).

The difficulty involved in determining the relationship between sea-level fluctuations and shoreline movement can be illustrated by computing the percentage of erosion that can be attributed to sea-level rise for the Avon reach of Hatteras Island and for all of Hatteras Island. The average predicted erosion rate, obtained from the three two-dimensional methods (−0.34 m/yr), is ≈113% of the observed average rate-of-change along the Avon reach when using the average linear regression rate of −0.30 m/yr and ≈49.3% of the observed average rate-of-change when using the average end point rate of −0.69 m/yr (for the period 1945–1986). Approximately 28% of the shoreline erosion for all of Hatteras Island can be attributed to sea-level rise (over the period 1945–1986) when comparing the predicted rate against the observed linear regression rate (−1.2 m/yr; for transects spaced at 50 m intervals) and ≈24% when using the average end point rate (−1.4 m/yr). EVERTS (1985), using a three-dimensional, control volume model, determined that sea-level rise accounts for 88% of the measured shoreline retreat rate.

As the above example illustrates, calculations of the percentage of erosion due to sea-level rise are clearly sensitive to model selection and to an accurate numerical representation of the model's terms. In addition, the spatial field over which the analysis is conducted, and the value selected for the observed rate-of-change, strongly influence this percentage. There is, however, qualitative evidence that a relationship exists between the sea-level curves and the shoreline data from Avon. LISLE (1983) was the first to relate short-term erosion rates (1952–1976) to sea-level rise for a 2.5 km reach near the Cape Hatteras Lighthouse. Comparing the Hampton Roads average annual sea-level curve to the shoreline data, LISLE (1983) postulated, by averaging discrete time intervals for 51 stations at 50 m intervals, that a 1966–1968 shoreline accretion pulse may have been due to a 1963 sea-level low period (Figure 5). This four year time lag between beach response and elevated water levels agrees with the conclusions of KOMAR *et al.* (1991). Similar results are also evident from more recent data. As mentioned above, the sea-level data, plotted in Figure 5, show distinct similarities. Examining the sea-level curves relative to the Avon shoreline response time series

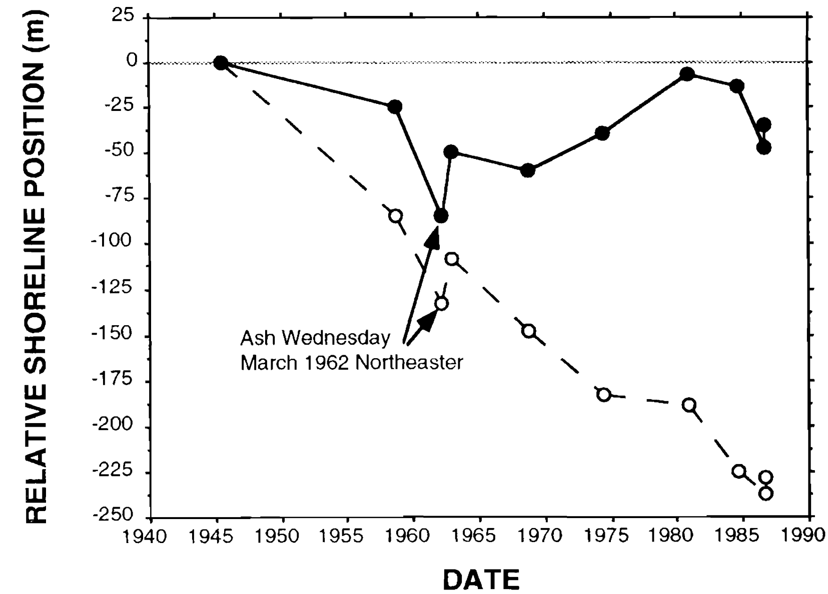

Figure 6. Shoreline data for two transects located on Pea Island, North Carolina. Open circles are data for a transect located adjacent to Oregon Inlet and solid circles are data for a transect located approximately 12 km south of Oregon Inlet. Shoreline movement for the latter transect does not appear to be affected by Oregon Inlet.

demonstrates that the continuing accretion trend corresponds to the post 1974 sea-level drop. This low sea-level stage approached 1963 low-levels. Following the 1974 low, sea level became highly variable, alternating between rises and falls, possibly influencing short-term shoreline response. In addition, using an eigenanalysis of tide gauge records between 1920 and 1983, BRAATZ and AUBREY (1987) showed an increase in the rate of relative sea-level rise (due to steric effects?) for the eastern United States *c.* 1934. This acceleration in RSL corresponds to the long-term erosion trends occurring at all the Avon transects between *c.* 1917 and *c.* 1970.

It should be noted that the sea-level history does not correspond to the shoreline history of each transect located along Hatteras Island. The island-wide impact of sea-level fluctuations may not be evident over the short-term due to more localized effects such as inlet dynamics; episodic, high-energy processes such as volumetric losses associated with storms; and/or anthropogenic factors. For example, volumetric sediment losses at Oregon Inlet may control shoreline dynamics for as far as 4 to 10 km to the south of the inlet (EVERTS, 1985; INMAN and DOLAN, 1989; DOLAN *et al.*, 1992). Figure 6 represents a typical transect eroding in response to inlet-related processes (located <1 km south of the inlet). The non-monotonic shoreline history of a transect located approximately 12 km south of the inlet-affected reach is typical of a non-inlet influenced shoreline (Figure 6). The shoreline response at this transect, therefore, may be more representative of shoreline changes associated with secular processes (*e.g.*, sea-level fluctuations) or short-term processes rather than those processes which are unique to distinct geomorphic features (*e.g.*, capes and inlets).

EVERTS and GIBSON (1983) state that the influence of a rising sea level and changes in shoreline positions relative to changes in the rate of sea-level rise are difficult to determine due to short and incomplete records and, therefore, cannot be used to predict absolute shoreline movement. Additional obstacles to using sea-level data to qualitatively assess historical shoreline changes result from: (1) using a nonlinear process (*e.g.*, sea-level history) to evaluate a nonlinear response (shoreline movement); (2) determining cause-effect relationships between S and R; and (3) comparing R predicted values to R observed values when the error range of those values is greater than the values themselves.

Oceanographic and Climatologic Regimes: Wind, Waves and Tides

North Carolina is a wave-dominated (microtidal) coast (HAYES, 1979). This is reflected geomorphologically by North Carolina's long and linear barrier system and by the predominance of storm washover deposits. The tides are semidiurnal with two approximately equal high tides and low tides occurring over a 12.4 hr cycle. Sound-side and ocean-side tidal circulation is dominated by the M_2 component of the astronomical tide, which produces a mean tidal range of approximately 0.3 m and 1.0 m, respectively. Spring tidal range is approximately 1.3 m. Although recent studies have shown that tide-induced currents can strongly influence fairweather sediment transport on the shoreface, the relative tidal contribution to the cross-shore flux over the short- and long-term is not well-known (*e.g.*, WRIGHT *et al.*, 1991).

Along the Outer Banks, shoreline movement is dominated by longshore and cross-shore sediment transport resulting from wave action. The mean wave height for the region is approximately 0.65 m, however, the wave climate is temporally and spatially variable (EVERTS *et al.*, 1983; INMAN and DOLAN, 1989; LEFFLER *et al.*, 1990). Approximately 8% to 9% of the deep water waves are greater than 1.5 m in height while approximately 0.25% of the deep water waves have heights greater than 3 m. The average deep water storm wave height is 2.5 m with a maximum storm wave height of 10.2 m. Deep water wave heights greater than 7.0 m, on the average, occur once every 25 years (INMAN and DOLAN, 1989). Wave direction is seasonally related to wind direction. Analysis of wave rose diagrams and wave data from gauge stations at Duck (1979–present; LEFFLER *et al.*, 1990) and Nags Head (1963–present; THOMPSON, 1977), North Carolina and from 20 years of hindcast data (1956–1975; JENSEN, 1983) indicates that 25% of all winds are from the northeast (from arctic and polar air masses), the predominant summer wave approach is southerly (tropical maritime air masses and cyclonic, low pressure systems) and the winter wave approach is from the northeast. These wind/wave patterns are responsible for producing a net southerly sediment flux of 590,000 m^3/yr for the reach located between Oregon Inlet and Cape Hatteras (INMAN and DOLAN, 1989).

The eastern seaboard is located on a leeward coast where long period waves, which produce substantial sediment (and shoreline) movement, result from storms rather than from swell. The highest waves at Avon are due to summer, tropical cyclones and anticyclones, and the more frequent and spatially expansive winter, extratropical cyclonic storms (northeasters).

Perhaps the most striking linkage of shoreline trends to wave climate is evident in the secular storm history of the North Carolina coast. High storm frequencies are observed at Hatteras due to its geographic location relative to the path of frontal storms, which propagate along westerlies, and to tropical cyclones. Using a hindcast data base consisting of 1349 extratropical storms for the period of 1942 to 1984, DOLAN *et al.* (1988) found no significant linear or monotonic long-term trend in extratropical storm frequency during these years. However, HAYDEN (1975, 1981), RESIO and HAYDEN (1975), DOLAN and HAYDEN, 1980), and DOLAN *et al.* (1988) have related shoreline changes to secular variations in cyclone frequency, magnitude and duration. These studies have shown that the mid-Atlantic storm climate has had the following characteristics:

(1) The length of the mid-Atlantic winter storm season increased between 1943 and the early to mid-1970's (HAYDEN, 1975; DOLAN *et al.*, 1988). Note the increased erosion occurring on all transects from the 1930's to the 1970's (Figure 3).
(2) The length of the winter storm season decreased beginning in the early to mid-1970's due to changes in storm tracks between 1973 and 1977 (HAYDEN, 1975; DOLAN *et al.*, 1988). Note the trend reversal occurring in the early 1970's (Figure 3).
(3) An anomalous decrease in storm-wave durations during the 1970's does not reflect a systematic long-term decline (HAYDEN, 1975; DOLAN *et al.*, 1988).
(4) Using a principal component analysis of mid-Atlantic cyclone frequencies over the period 1885–1978, RESIO and HAYDEN (1975) and HAYDEN (1981) provided evidence for a significant trend reversal in large-scale circulation patterns resulting from an increase in high-latitude, anticyclone blocking. These circulation patterns corresponded to a decreased cyclone frequency over the mid-Atlantic from 1885 to 1925 followed by an increase in frequency peaking in the 1960's. A 15% increase in extratropical storms from the

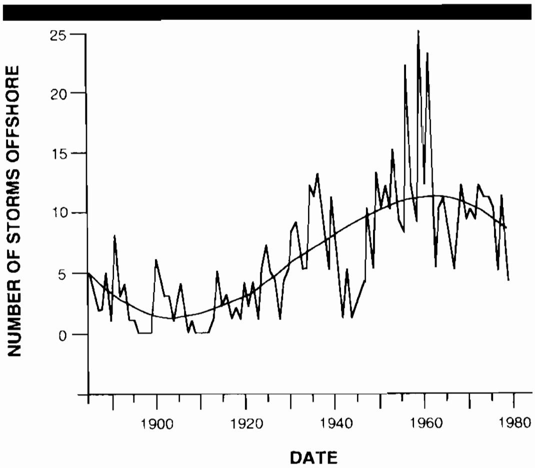

Figure 7. Frequency of offshore storms since 1885 in the Cape Hatteras region. Note the possible 100-year cyclicity involving increasing storm activity during the late 1900's to the early 1960's (From DOLAN and HAYDEN, 1980).

1920's to the 1960's is part of a secular variation of a longer time scale whose periodicity is unknown. The increase in storminess from the 1920's to 1960's corresponds to the erosion trend reversal (second temporal trend) on the Hatteras transects, while the decrease in the 1960's and years following correlates well with the accretion trends at Avon and the decline in erosion at other localities along Hatteras Island.

(5) Frequency data displaying the number of storms passing through a 2.5° latitude by 5° longitude cell off the North Carolina coast further demonstrate the secular variation of extratropical storms over the 93 year period of record and the increased storminess from the 1920's into the 1960's (Figure 7; DOLAN and HAYDEN, 1980).

Tropical storms and hurricanes occur on a much less frequent basis and impact a smaller reach of coast than extratropical storms. None the less, SIMPSON and RIEHL (1981) demonstrated a trend in tropical storm and hurricane occurrence which is similar to that of northeasters: A low frequency period occurred at the end of the 1920's with below average frequencies from 1895 to 1930. Beginning in the 1930's, hurricane frequency increased significantly until the 1970's. The lowest hurricane incidence level of this century occurred in the mid-1970's.

Although these studies have shown that a change in the structure and occurrence of mid-Atlantic coastal storms has occurred during the last four decades, continuation of the long-term decline in storm duration and increase in frequency cannot be predicted with any certainty (DOLAN *et al.*, 1988). Thus, although these studies and others (*e.g.*, ALLEN, 1981; BRYANT, 1988) have linked shoreline trends and wave climate, forecasts of shoreline locations are hampered by an inability to accurately assess future wave climate trends (EVERTS and GIBSON, 1983).

Controls on Sediment Supply

Shoreline movement can result from volumetric changes in the sediment budget. Volumetric losses or gains of sediment within a reach (or control volume) are directly linked to changes in sediment supply, delivery, and storage capacity over time or to spatial variations in the longshore transport rates (*e.g.*, ALLEN, 1981). These changes may be driven by natural or anthropogenic factors and occur over a wide range of temporal scales. The challenge in linking the processes with their responses lies in determining which changes are permanent and which are temporary (MORTON, 1991). Temporary changes in the sediment budget can account for significant variability (noise) in long-term shoreline movement.

Natural causes of regional changes in the longshore sediment budget include the downdrift propagation of the sediment in the form of distinct coastal geomorphic features (*e.g.*, rhythmic features) and sand "bulges", *i.e.*, accretion and erosion waves (MORTON, 1979; ALLEN, 1981; DOLAN and HAYDEN, 1983; INMAN, 1987). These features result from an episodic supply of sediment to the littoral zone (*e.g.*, floods, ephemeral rivers, dune erosion) or from changes in the structural organization of the oceanographic processes controlling their formation (*e.g.*, edge waves).

As an example of longshore temporal variations in sediment transport, INMAN (1987) showed that the amplitude of an accretion and erosion wave attenuates downdrift by diffusion, but the travel velocity of the wave increases. Interestingly, accretion waves retard longshore transport and result in localized downdrift erosion. For the California coast, INMAN (1987) observed average accretion wave speed in the near field (defined as the region of large amplitude waves near the source) between 0.5 and 1.5 km/yr and far field rates of 2 to 4 km/yr. For the North Carolina coast, DOLAN (1970) observed "sand waves", with wavelengths of 150 to 915 m, migrating at rates up to 2.2 km/yr. Comparing 1937 and 1976 shorezone profile data along a 71 km reach on the Outer Banks, FISHER *et al.* (1984) found that the rhythmic forms of different frequencies can migrate along the coast at different rates or migrate at similar rates. In general, however, the rhythmic forms migrate as much as one quarter wavelength in four decades (FISHER *et al.*, 1984). Using the reported 17.8 km form wavelength, migration occurs at rates ranging from 100 m/yr to 150 m/yr (corroborated by visual examination of aerial photographs).

The volume of sediment associated with the intermittent and episodic migration of sedimentary landforms is difficult to quantify with available data. One significant effect of migrating features or sediment "pulsing" in a longshore direction, however, is to distort (add noise to) the actual shoreline position. For example, a shoreline located at the horn of beach cusp will mark an apparent seaward shoreline position, while the shoreline location at a trough will show an apparent landward position. PAINE and MORTON (1989) determined that variability in the shoreline's position due to subharmonic rhythmic features along the Texas coast ranges from 10 to 60 m.

Dune sands provide a potential, additional source of sediment to the Avon reach (STAUBLE, 1979; LEATHERMAN, 1979; and FISHER *et al.*, 1984). In the 1930's, the National Park Service constructed a twin dune ridge system from the Virginia-North Carolina state border south to Ocracoke Island, North Carolina, for the purpose of providing a barrier to storm surge and washover deposits. DOLAN (1972), GODFREY and GODFREY (1973) and HAYDEN *et al.* (1980) documented the beach changes which resulted from the dune stabilization project. Post stabilization changes included a winnowing (eroding) of the active beach as a result of compressed wave energy dissipation into a narrower shorezone as well as a steepening of shorezone profiles. The National Park Service maintained the dunes until a 1972 policy reversal. Beginning in 1972, the artificial dunes were allowed to return to their natural state. FISHER *et al.* (1984) concluded that the shifting spatial positioning of the topographic lows and highs, due to the erosion and aggradation of dunes during this period of reestablishing the natural dune system, may be a primary process which dictates erosion-prone versus erosion-safe areas. Therefore, prior to 1972, an important source of beachface sediment (fine- to medium-sized sand in the range 0.20 mm to 0.40 mm; DOLAN 1972) was temporarily (and artificially) placed in storage and after 1972 an indeterminable quantity of sand became available to the beachface.

The trend reversal at Avon during 1968–1972, from relatively high rates of erosion to low rates of accretion, must coincide with a volumetric sediment gain to the beach system. To date, however, the dunes fronting the town of Avon remain intact and the trend change in the Avon data may have occurred prior to the decision to allow the dunes to return to their natural state. Moreover, no difference in the magnitude or duration of trends is evident in the data for base map, transect 20-1 which is located south of Avon in a region of completely eroded dunes (Figure 3). Thus, shoreline movement at Avon does not appear to be responding to an influx of dune source material from the immediate vicinity.

Updrift dune erosion may possibly explain the sediment influx to the Avon beaches. DEKIMPE (1987) delineated four reaches which displayed dune erosion in 1986 from Oregon Inlet to Cape Hatteras. Three of the eroded dune reaches are located north of Avon and the southern reach extends from Cape Hatteras to 3 km south of the Avon fishing pier (base map 20, transect 16). Northern sediment sources are particularly important to the Avon beaches due to the net southerly transport direction (INMAN and DOLAN, 1989). In 1986, a 3.2 km segment of the dune ridge system, approximately 8 km north of Avon, was partly to completely eroded. Assuming a triangular dune shape, an average dune height of 3 m, and a width of 60 m (DEKIMPE, 1987), approximately 288,000 m^3 of sand-sized sediment could be available to nourish the beach and/or nearshore zone. Visual observation of the shoreline data north of base map 21 corroborates this finding of a continuous accretionary pulse in the reach from base map 21 up to the northern end of the eroded dunes (base map 24, transect 30) over the time span from 1968 to 1986. DEKIMPE *et al.* (1991) predicted that by the year 2000, 22% of the dunes will be eroded along the Outer Banks and, over the next century, 70% of the artificial dunes will return to their natural state, thereby providing a significant quantity of sediment to the Outer Banks.

Although net southerly transport dominates the Hatteras Littoral Cell (INMAN and DOLAN, 1989), a beach nourishment project south of the study area (intended to protect the Cape Hatteras Lighthouse) may have provided a source of sediment for the Avon beachface due to annual variations in the longshore transport direction. A 2.1 km-long reach received 238,000 m^3 of sand-sized sediment in 1966, 150,000 m^3 in 1970–1971, and more than 1,000,000 m^3 in 1973 (LISLE and DOLAN, 1984). During the 1973 replenishment project the northern-most discharge point occurred at a location approximately 2.7 km north of the Cape Hatteras Lighthouse and approximately 6.0 km south of base map 20. Visual examination of shoreline data south of base map 20 (*i.e.*, 17–20) indicates that, although the 1973 project produced a 91.5 m accretional bulge, the effects of nourishment were localized and, north of base map 18, negligible.

Additional processes which control differential longshore transport rates include variations in wave refraction patterns and changes in cross-shore sediment fluxes (*e.g.*, DEAN, 1987; WRIGHT *et al.*, 1991). A significant factor controlling wave approach patterns and thus, cross-shore and longshore transport in the Avon region is the complex bathymetry in the nearshore zone (Figure 1). A series of prominent northeast-southwest trending shoreface-connected ridges, approximately 10 m in height and in water depths of 7 m, appears to control a shoreline response. EVERTS *et al.* (1983) found a relationship between the alongshore variation in shoreline change and these shoreface connected ridges. EVERTS *et al.* (1983) showed that reaches north of the shoreface and ridge system intersection were consistently erosion-dominated and reaches south of the intersection were accretion-dominated. Although the relationships seemed consistent, EVERTS *et al.* (1983) found no physical cause for these differences. Although the origin of similar features along the mid-Atlantic coast has been debated (*e.g.*, DOLAN *et al.*, 1979; SWIFT, 1980), RIGGS (in press) provides evidence that many of the ridges along the Outer Banks are nearshore outcrops containing a paleo-nucleus of Pleistocene-age material. RIGGS (in press) postulates that these linear relict mud-mounds (veneered with sand-sized sediment) are nearshore promontories which geologically control the geometry (prominence) of the shoreline. Thus, the hardground outcrops are more erosion-resistant than the erosion-prone lateral facies.

The shoreline data from Avon do not completely concur with the results of EVERTS *et al.* (1983). First, contrary to the description of EVERTS *et al.* (1983), the intersection of the ridge system and shoreline cannot be represented by a single geographic coordinate. For example, Kinekeet Shoals comprise three distinct ridges with slightly different orientations (30°, 45°, and 60° north to south). The alongshore distance separating the intersection of the shoreface and northernmost ridge and the intersection of the shoreface and southernmost ridge is approximately 5.5 km. Thus, the assertion of EVERTS *et al.* (1983) that Kinekeet Shoals intersects the shoreface at 35°23′N latitude and the shoreline north and south of this intersection is retreating and accreting, respectively is an oversimplification. The data in this study (base map 20, transect 1 to base map 21, transect 70) span from approximately 35°22′ southward to 35°18′35″N latitude (≈7 km). According to EVERTS *et al.* (1983) all transects south of 35°23′N latitude should show prograding shorelines. EVERTS and GIBSON (1983) state, however, that shoreline changes associated with shoreface connected ridges are only predictable in a broad relative sense, but not on a scale of several kilometers or less. In addition, EVERTS and GIBSON (1983) found that the shoreline behavior associated with Kinekeet Shoals is unique compared to the other ridge systems in the Hatteras Littoral Cell. Thus, the role of the ridges in controlling shoreline trends continues to be ambiguous and requires further study.

Wave refraction analyses can be used to determine the distribution of wave energy along the coast. Since the path of a wave ray is dependent upon bathymetry, however, it follows that the angle of wave approach for regions consisting of complex ridge and bar morphologies is highly variable. Because refraction diagrams are limited by the accuracy of depth data, by ranges in tide levels and wave periods, and by the assumptions of linear wave theory and refraction models, wave refraction models may be of limited value in regions consisting of complex bathymetry. For example, the wave refraction models of FISHER *et al.* (1975) and GOLDSMITH (1975) show that the area north of Avon is one of wave ray convergence. Interestingly, however, the expected trend towards erosion is not consistently observed over the long-term in this area (*e.g.*, base map 21, transects 60 and 70). In contrast, the region south of Kinekeet Shoals does show a relatively long-term trend towards erosion, which is consistent with the expected concentration of wave energy downwave from a topographic high (GOLDSMITH, 1975).

Other Considerations

The influence of engineering and hard structures on shoreline movement is relatively well-known (*e.g.*, KOMAR, 1983; among many). Except for a terminal groin at the southern end of Oregon Inlet and three small groins fronting the Cape Hatteras Lighthouse, the existing structures along Hatteras Island consist of two relatively small fishing piers. One pier is located approximately 30 km north of the study area near Rodanthe, North Carolina, and the other is located at Avon near base map 21, transect 16.

The 260 m long fishing pier at Avon was constructed in 1964. Examination of sequential aerial photographs reveals that the impact of the pier is local. This impact is evidenced by a 100 m long and 20 m wide shoreline bulge which diminishes with distance away from the pier. Approximately 300 m downdrift of the pier, the shoreline is eroding linearly, over the long-term, at a rate of approximately −0.4 m/yr (base map 21, transect 10; FENSTER *et al.*, 1993). Although the pier's presence may influence shoreline movement at this location, the data suggest that the shoreline history at transect 21-10, prior to the pier's construction, is anomalous. Sequential shoreline position data at this transect reveal an erosional period prior to *c.* 1917 (Figure 3). During this same time, transects at least 3 km north and 2.5 km south of this site recorded accretion. Furthermore, the anomalous shoreline behavior at this location is witnessed by an episode of minor accretion followed by moderate erosion occurring concurrently with a period of severe erosion at all the other transects examined in this study (*c.* 1917–1968; Figure 3). The observed trends at transect 21-10, therefore, cannot be explained using existing oceanographic and geologic data.

Other factors which possibly contribute to the region's shoreline dynamics, but for which no quantitative data are available, include fluctuations in the ground water table (CLARKE and ELIOT, 1987), wind transport, beach deflation, bar morphology, rip current cells, and/or variations in beach states (*e.g.*, dissipative or reflective). Finally, although the role of historical inlets must be considered, the only historical inlet in existence near Avon during the study period was Cape Inlet, located approximately 5 km south of Avon. According to available charts and aerial photographs, Cape Inlet was open prior to 1657 (INMAN and DOLAN, 1989) and for 10 months following the 1962 Ash Wednesday storm (The U.S. Army Corps of Engineers artificially closed the most recent inlet in January 1963.) Thus, inlet-related processes played a minor role in the relatively recent (≈134 years) shoreline history of the Avon reach.

DISCUSSION AND CONCLUSIONS

Temporally sparse, error-prone response measurements, and the synergistic nature of those conditions which produce shoreline changes effectively constrain our ability to link the shoreline response patterns on most of the world's coastlines (including the Outer Banks) with oceanographic processes. To complicate matters still further, the coastal processes occur over a wide range of nested, temporal scales (Figure 2) resulting in a long-term signal and short-term variability. Although the impact of an individual (usually short-term) process on sediment transport can be modelled (*e.g.*, WRIGHT *et al.*, 1991), the methods needed to quantify the cumulative effect of synergistic, noisy processes have not been developed to date.

Given the above limitations, the trend analysis technique developed by FENSTER *et al.* (1993), which used the time series data for the Avon reach along the Outer Banks, has demonstrated that, north and south of base map 21 transect 10, two prominent changes in the long-term trend have occurred. The earliest change most likely occurred prior to 1917 and the most recent trend reversal occurred between 1968 and 1972. At best, the recent trend towards shoreline accretion can be related qualitatively to the cumulative effect of three phenomena: (1) a fluctuating sea level with several short-term low stands, (2) a decrease in storminess (*i.e.*, length of winter storm season, storm wave durations, and storm frequency), and (3) an influx of sand-sized (probably dune) sediment (Figures 3, 5, and 7). The long-term erosion interval between the early and late reversal may be attributed to: (1) an increase in extratropical storm activity during the period from the 1920's to the 1960's, (2) a steadily rising sea level, and (3) the influence of an artificial dune ridge system. No other quantitative, time series data exist for this area regarding other process-response patterns (i.e., changes in groundwater levels, alongshore propagation of sand bulges, longshore variations in sediment transport, the impact of offshore connected ridges on sediment transport processes and wave refraction patterns and/or variations in sediment supply). Existing wave refraction analyses in this region are limited by the accuracy of bathymetric data, by ranges in tide levels and wave periods, and by the assumptions of linear wave theory and refraction models; thus, a clear relationship between areas of wave ray convergence and beach erosion is not apparent. The shoreline in the vicinity of base map 21, transect 70, however, may be less erosion-prone than transects south of this site (and as much as 2 km north) due to the location of Kinekeet Shoals and the coincident reduction in wave energy.

One conclusion of this study is that both the direction and magnitude of shoreline movement in time are spatially variable; a shoreline's response to a set of processes is not spatially persistent. This indicates that, over the time scales covering historical map and aerial photographic sample data, the signal in shoreline movement (associated with long-term processes) can be obscured (KOMAR *et al.*, 1991). We agree, therefore, with KOMAR *et al.* (1991), who conclude that application of the Bruun Rule, or a modification thereof, is error-prone and of limited value when used on a site-specific basis. BRUUN (1988) agrees that the application of the rule is limited if the shore is not located in a neutral area with respect to littoral drift. In addition to confirming the observations of KOMAR *et al.* (1991) regarding the Bruun Rule, we cannot substantiate EVERTS' (1985) conclusion that long-term sea-level rise is the predominant cause of shoreline erosion over the 134 year study period along the Outer Banks. Thus, we conclude that the observed responses at Avon result from relatively local, shorter-term processes which alter the sediment budget.

Comprehensive models and quantitative correlations for coupling physical processes to shoreline response to not exist. Our recent analytical research (DOLAN *et al.*, 1991, 1992; FENSTER *et al.*, 1993) has given us the opportunity to review the available methods for understanding shoreline response do not exist. Our recent analytical specific basis. Our inability to distinguish and to quantify each individual factor causing alterations to the sediment budget, as well as to quantify the cumulative effect of these processes in the Avon area, is characteristic to the study of most open ocean coasts. Future research directed toward quantifying and modelling the time-dependent process-response coastal system and the variability due to short-term processes is required in order to improve the reliability of shoreline rate-of-change values and our ability to develop accurate theoretical models.

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