



Coastal Erosion and Solutions

A Primer

2nd Edition



COASTAL SCIENCE & ENGINEERING

Coastal Erosion and Solutions

A Primer 2nd Edition

ACKNOWLEDGMENTS

Coastal Erosion and Solutions – A Primer was first offered in 2003, drawing from studies and beach restoration projects completed by Coastal Science & Engineering since 1984. It benefitted particularly from research by former colleagues of the author at the University of South Carolina's Coastal Research Division (Professor Miles O. Hayes, Director) and many professionals, some of whom are listed in the bibliography. This edition is dedicated to Dr. Nick Kraus who has been a tireless champion of our profession.

This second edition is offered partly to update some new findings but, more importantly, to provide more examples of practical solutions. After designing over 35 beach restoration projects involving nearly 25 million cubic yards, the CSE team has developed a keen appreciation of the need to work with, rather than against, nature. Not all solutions are long-lasting when it comes to coastal erosion. Yet, the number and geographic extent of beach restoration successes are large and growing, driven by improved understanding and engineering as well as by burgeoning demand for lasting solutions.

We especially thank the municipalities and property owner associations at the coast who have supported our designs and have given us the privilege of measuring their beaches year after year. Without such data collection sponsored by Myrtle Beach, Seabrook Island, Bogue Banks, Nags Head, Kiawah Island, Isle of Palms, Hunting Island and Edisto Beach—to name some of our project sites, our understanding of coastal erosion at mesoscales would still be in its infancy.

CONVERSION FACTORS

| Multiply | By Approx | To Obtain |
|--|-----------|--------------------------------|
| Millimeters (mm) | 0.04 | Inches |
| Centimeters (cm) | 0.4 | Inches |
| Meters (m) | 3.3 | Feet |
| Kilometers (km) | 0.6 | Miles |
| Square meters (m ²) | 10 | Square feet (ft ²) |
| Cubic meters (m ³) | 35 | Cubic feet (ft ³) |
| Cubic meters (m ³) | 1.3 | Cubic yards (cy) |
| Cubic meters per meter (m ³ /m) | 0.4 | Cubic yards per foot (cy/ft) |

Note on Units: An early version of this booklet was geared for an international audience and, therefore, SI metric units were used. We give preference to metric units herein, but frequently provide English-unit equivalents to assist the reader. Considering that coastal erosion at mesoscales necessarily entails quantification, it is useful to speak the language of arithmetic in metric as well as English units.

A related paper, entitled *Coastal Erosion and Defense Against Erosion*, is available by subscription in the web-based Encyclopedia of Life Support Systems (www.eolss.net), published by UNESCO (2003, Oxford, UK)

Cover Photos: *Front* – Seabrook Island and Captain Sams Inlet, South Carolina (looking north) at low tide in April 2010 after restoration of the beach by inlet relocation and forced sand bypassing. *Back* – The same area at low tide in 1982 before inlet relocation. Note armored shoreline protecting development.

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PREFACE

Coastal erosion is often thought of as inevitable. The forces of winds, waves, and currents on the shore are uncontrollable. Tiny particles, like the sands that make up the great recreational beaches of the world, will move inexorably from place to place. As sediment moves, so does the coastline. But viewed in human time scales of decades to centuries, many beaches are moving imperceptibly. After all, they have had thousands of years to evolve into forms that are nearly in balance with the local wave and tide conditions. They may erode during storms, but often rebuild naturally in a continuing cycle. Human activities have exacerbated erosion in many areas, but so have large-scale phenomena like channel avulsions or natural openings of inlets and hurricanes or tsunamis.

This primer describes some of the causes of coastal erosion and tries to put into perspective their scales and consequences. There are no uniform causes, just as there are no uniform solutions. Erosion tends to be site-specific. Yet, with careful observation and measurement, a particular problem can be placed in context and draw from the experience of similar sites. Given the variety of the world's coastlines, many other "signatures of erosion" beyond those mentioned here are at work. The attraction for scientists seeking to understand these signatures is the same as the casual tourist's—the ever-changing image of the shore.



Hunting Island (SC) at low tide in April 2010 after eight nourishment projects (1968–2006) and construction of six groins (2007) designed to retain sand along segments of the 6.5-kilometer-long (4 miles) barrier island. Few sites have posed greater challenges for beach restoration, given Hunting Island's historical erosion rate of ~8 meters (~25 feet) per year. [See Traynum et al (2010) for a summary of shore-protection measures and the semi-soft solution CSE designed for the island.]

INTRODUCTION

All sedimentary coasts tend to erode at one time or another. This basic tenet of coastal science reflects the complex interactions that occur where air, sea, and land come together. Wherever shorelines are composed of discrete grains of sediment, the processes of winds, waves, currents, and changing water levels combine to mobilize the particles and move them around by varying degrees.

This primer describes the physical processes and primary causes of coastal erosion and presents some commonly applied erosion defenses. To many observers, erosion is inevitable and a problem only because of human development at the coast (Bush et al 1996). Yet, on closer examination, the problem and solutions have many facets. Not all coasts are alike, nor are they all eroding. Each has its own set of characteristics, including specific geologic history, sediment type, exposure to waves and tides, and relationship to everyday use by humans.

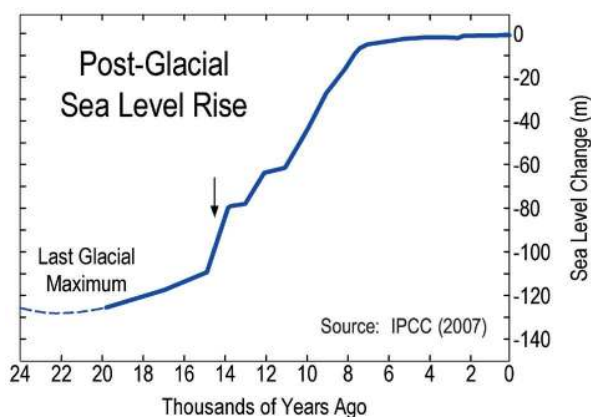
As long as people are drawn to the sea, the coast will remain a special place, having to adjust to storms, rising sea levels, and encroaching development. The question is not so much whether the coast is eroding, but how we can live with it and properly accommodate its changing shoreline conditions.

SEA LEVEL VARIATIONS

Coasts evolve in relation to their tectonic setting, sediment supply, wind and wave climate, and water-level changes. Over long geologic time scales [>1 million (10^6) years], plate tectonics fundamentally control the evolution of the coast (Inman and Nordstrom 1971). Where the earth's plates are colliding, relief at the coast tends to be steep.

This results in narrow coastal plains and continental shelves. On the other hand, where the earth's plates are separating or "trailing away" from each other, relief tends to be gentle, leaving wide continental shelves and coastal plains. The Pacific and Atlantic coasts of the United States are examples of a "collision" coast and a "trailing-edge" coast (respectively). Most barrier island and estuarine shorelines are situated along trailing-edge coasts.

At much shorter geologic time scales ($\sim 100,000$ or 10^5 years), sea-level movements control the position of the coast. Examination of sediments and fossils confirms that sea level has fluctuated by at least 100 meters (m) over the past 100,000 years (Fig 1). Cycles of glaciation (the most important sea-level position factor) release or remove huge volumes of water from the world ocean, causing seas to rise or fall across the coastal plain and continental shelf. Today, sea level is closer to its highest possible level than its lowest recorded level because the earth is between glacial periods. Most researchers agree the last period when sea level was close to its present position was around 120,000 years ago. Interestingly, as the graphic in Figure 1 shows, sea level has not spent much time near today's level.



Late Pleistocene and Holocene Sea-level Curve

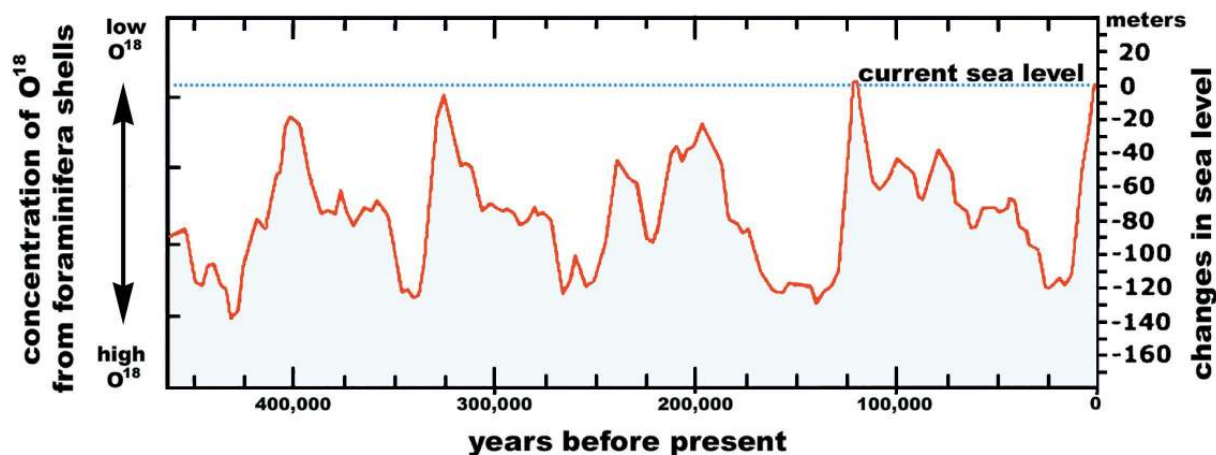


FIGURE 1. Sea-level curves inferred from oxygen isotope ratios showing peaks correlating with interglacial periods and the eccentricity of the earth's orbit around the sun which has a period of $\sim 100,000$ years. [Source: Imbrie & Imbrie 1979]

At historical time scales (~10,000 or 10^4 years), the magnitude of sea-level change has been of the order 10 meters (m) (Fairbridge 1961). Carbon-14 dating confirms the last glacial stage (“Wisconsin”) ended with a global warming trend around 20,000 years before present (BP). Between 20,000 BP and 5,000 BP, melting ice sheets produced a rapid rise in sea level (Shepard 1963). Present sea level was nearly reached 5,000 years ago, at which time its rate of rise slowed markedly. Since then, sea level has oscillated over a narrow range with some periods experiencing a lowering and others a rising ocean.

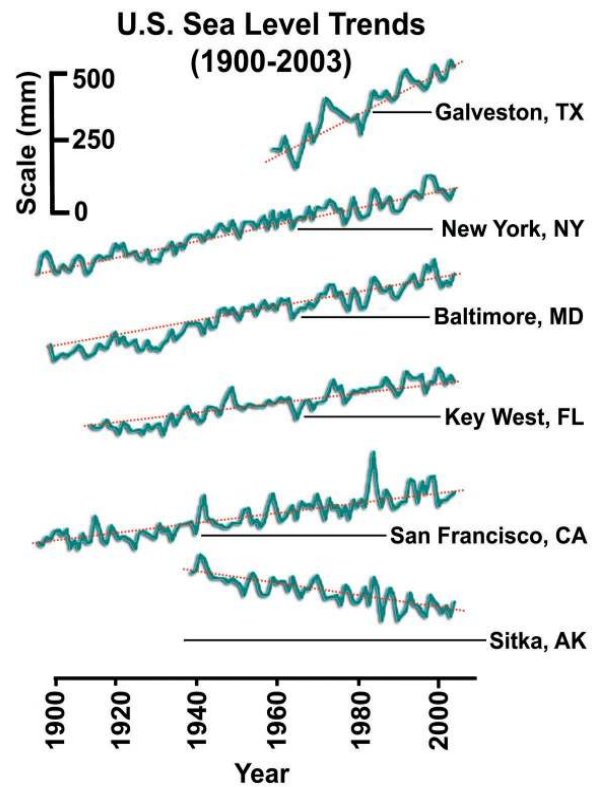


FIGURE 2. Chesapeake Bay, a drowned river estuary, and the outer coast of the Delmarva Peninsula (US Atlantic Coast), a coastal plain shoreline that formed within the past 5,000 years after sea level nearly reached its present position. [Image Source: ESRI i-cubed 2010]

Our present coastline and its diversity of features (including river deltas, barrier islands, lagoons, and estuaries) all formed in the past few thousand years (Fig 2). In geologic terms, then, the coast is exceedingly young and relatively ephemeral (Schubel 1972). It may persist in more or less its present shape for several thousand more years, or it may migrate seaward or landward if the global climate destabilizes, triggering a new glacial stage or a warmer cycle. Glaciation removes volume from the ocean and leads to lower sea levels. Global warming, of course, does the opposite, melting the ice caps and raising sea levels.

In the last decades of the 20th century, attention was focused on global warming because of evidence that the industrial age and human activities are producing a measurable impact on climate (Barth and Titus 1984, Kellogg 1988). The burning of fossil fuels and other anthropogenic effects have increased carbon dioxide (CO_2) concentrations in the atmosphere and, through the greenhouse effect, contributed to global warming of the order 0.5 degrees Celsius ($^{\circ}C$) in the 20th century. Evidence suggests that the mean global temperature has fluctuated no more than $\sim 1^{\circ}C$ over the past 1000 years and about $5^{\circ}C$ in the past 25,000 years (Hansen et al 1984). Whether or not mankind has been responsible for recent global warming, climate change models (IPCC 2007) now project a probable $\sim 2^{\circ}C$ increase in global temperatures over the 21st century. Warming of this magnitude could lead to sea levels 18 centimeters (cm) to 59 cm (~ 0.6 – 2 feet) higher than 2000 levels by 2100 (IPCC 2007). This compares with a global rise of 10–12 cm during the 20th century (NRC 1987).

Sea level is such a fundamental parameter that coastal erosion tends to be closely linked to any rising or falling trend. The degree to which a changing sea level will alter the shoreline depends on the slope of the land at the coast. Gently sloping, low coastal plains (such as the shoreline around the Brahma-Putra River in Bangladesh) will tend to fluctuate much more than high-relief coasts such as Baha, California. However, a close examination of world shorelines reveals great complexity in their response to



changing sea levels (NRC 1987). Some coastlines retreated much further than expected during the 20th century (in the face of a 10–12 cm global sea-level rise), whereas others actually grew seaward.

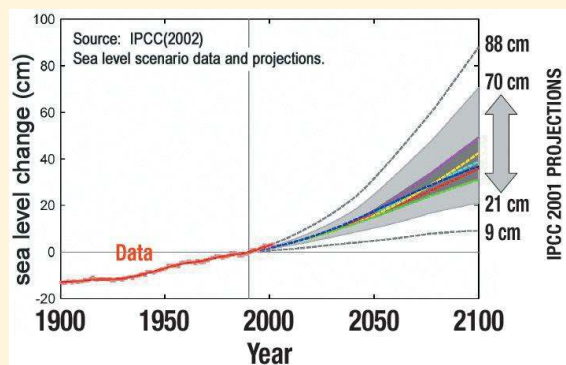
Some of the differences in shoreline erosion rates are due to varying degrees of land subsidence. The mouth of the Mississippi River accumulates muddy sediments that compress and contract in volume over time. Combined with the extraction of gas and oil deep below ground, sediments that form the Louisiana coast are sinking at a rate many times the global rate of sea-level rise. Thus, the “local” or “relative” rate of sea-level rise for Louisiana has been of the order 1 m per century (ie – ~10 times the global, or “eustatic,” rate of sea-level rise). The rate for South Carolina, which has less muddy sediment delivered to the coast and no oil or gas extraction, has been ~24 cm per century (ie – approximately double the global rate). The rates of change vary due to the local effects of subsidence added to whatever average global change

occurred in that time frame. In these examples, Galveston (TX) shows the highest rate of local sea-level rise (~6.5 mm/yr); New York City’s rate has been ~2.8 mm/yr. Note the result for Sitka, Alaska, shows a drop in local sea level, which in that case, reflects tectonic activity and “rebound” of the land following deglaciation over the past 20,000 years. Similarly, the coast of Maine is rising, giving the impression to a very old and patient observer on land that the ocean is receding.

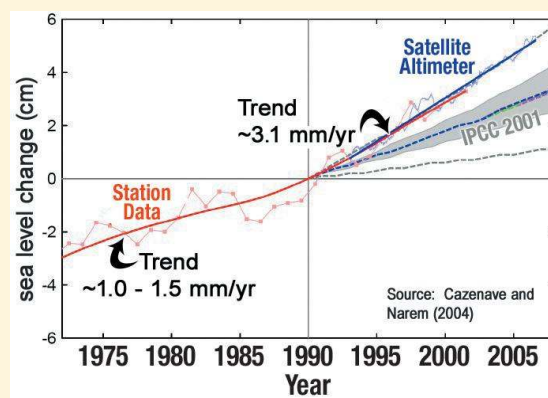
Maine’s good fortune, with respect to local sea-level trends, combined with Louisiana’s bad situation offer lessons for the future. The response of the coast to future sea-level rise will not be uniform from place to place. Relative sea-level rise is but one factor accounting for differences in shoreline movement. But before outlining why such variability occurs, it is necessary to establish a common frame of reference because coastal erosion is a time-dependent process.

Recent Sea-Level Rise Projections

Since the 1980s, there have been hundreds of studies related to global warming, its cause, its prediction, and potential impacts. Distinguished panels convened by the National Academy of Sciences (NRC 1987) and the Intergovernmental Panel on Climate Change (IPCC 2002, 2007) have been assimilating extraordinary data sets and modeling at global scales to project future trends in temperature, weather patterns, and sea level. There is uncertainty with all projections, yet a consensus has emerged that increased warming will lead to higher sea levels. The question is how much higher over what time frame?



The accompanying graphs show sea-level rise scenarios that are at the heart of the current (2010) debate. Climate models presently project a probable ~2°C rise in temperature over the 21st century. This doubling (or more) in the temperature rise compared with the last century is expected to correlate with sea-level rise. Using much refined analyses since the pioneering models of the 1980s, the IPCC (2007) projects global sea-level rise in the range 18-59 cm by approximately 2100. This compares to a worldwide increase of 10-12 cm during the 20th century. (Keep in mind, this is a eustatic or global average and does not include local effects of subsidence.) While the most recent IPCC (2007) projection has narrowed the scenario range compared with those of the 1980s (NRC 1987), many scientists believe it underestimates because it does not account for “sudden” thawing of glaciers. Even the most sophisticated models are unable to account for “tipping points” in global temperatures which would presumably trigger mass melting in shorter



time periods. Thus, the IPCC (2007) “high” scenario, sea-level rise of 59 cm (~2 ft) in the next 80-90 years may turn out to be optimistic. Some advocates of higher projections suggest a “5-7 feet” rise by 2100 should be anticipated.

Yet, counterbalancing those who argue for higher sea-level rise scenarios are other scientists who point to limited evidence of accelerated sea-level rise (Houston and Dean 2010). Tide-gauge records for certain stations during the latter half of the 20th century show a steady, but not clearly increasing, sea-level rise. Some of the interpretations of sea-level data are complicated by use of different time periods and the introduction of satellite telemetry, a more recent technology which is a fundamentally different measure of water levels compared with the global network of tide gauges. Some data can be used to calculate a 1–1.5 millimeter per year (mm/yr) rise during portions of the 20th century and a ~3 mm/yr rise (using satellite telemetry) between 1993 and 2003. The latter rate is equivalent to 30 cm (~1 ft) per century. Notwithstanding the uncertainty of comparing data obtained by different methods, the measured trends presently fall toward the lower scenario range of projections. Clearly, to achieve a 59-cm rise (or higher) by 2100, the rate of sea-level rise must increase dramatically, and the evidence for the higher scenarios will have to manifest soon.

While scientists continue to debate the evidence and rates of accelerated sea-level rise, we offer some practical guidance for coastal communities under Coastal Erosion Defenses.



FRAMES OF REFERENCE

Scales of coastal change generally increase the longer one looks backward or forward in time. Over long geologic time scales, tectonic activity, volcanism, glaciation, and even biogenic processes (such as reef building) will fundamentally control the shape and position of the coast. Where broad continental shelves occur (such as the U.S. East Coast), shoreline position will fluctuate 50–100 kilometers (km) or more in the cross-shore direction. However, geologic time scales are well beyond the period of concern for global habitation and planning. At the opposite extreme from geologic time scales are instantaneous events such as those associated with a single breaking wave at the shoreline or perhaps landfall of a major storm. Certainly, it is important to understand these microscale processes but, by their very nature, it is difficult to extrapolate from isolated events. Human time scales are generally of most concern because our investments parallel our life spans. These middle, or “meso,” time scales are a practical frame of reference because they can be linked to recorded history and personal experience. They allow averaging of many microscale events while avoiding the unpredictability and interminable pace of most tectonic activity.

For the remainder of this primer, coastal erosion will be considered in the context of a mesoscale frame of reference; that is, shoreline fluctuations experienced over decades to centuries. The U.S. government establishes minimum building elevations based on the 100-year flood (defined as the expected flood elevation for a particular coastal or riverine site that has a 1 percent probability of occurrence in any given year). Some states, such as South Carolina, base coastal development on “40-year” setback lines (SCCC 1991). In that case, development must be placed landward of the anticipated maximum point of erosion over the next ~40 years, using historical shoreline data for the previous ~40 years or longer.

Setback lines are political jurisdiction lines that establish the seaward-most limit for human development along a particular coast based on any number of criteria depending on the locality, governing authority, and available shoreline data. They may prescribe an arbitrary distance landward of the present shoreline or be linked to site-specific erosion rates. Although there is nothing unique about a particular time period, choice of one provides for consistency in the application of laws and building codes. Longer planning periods are often desirable but too costly to implement compared with the economy at risk. However, in places like The Netherlands, where large populations and much of a country’s wealth is at stake along the coast, planning horizons should expand well beyond 100 years.

Conflicting views regarding coastal erosion stem largely from different frames of reference. Since the late 1970s,

there has been heated debate among geologists, coastal engineers, environmentalists, and the media over the problem of coastal erosion and what to do about it (Houston 1990, Pilkey 1990). Such debates can only be tempered if common time frames and scales are adopted. In the meantime, those who live, work, and play at the coast need to understand that coastal erosion is site specific and so must be accommodated based on local conditions (NRC 1995).

COASTAL PROCESSES AND EROSION

Unconsolidated sediments at the coast move in response to winds, waves, currents, and changing water levels. As we saw in the previous section, global effects control sea level. However, at mesoscales, the magnitude of sea level change is relatively small, on the order of 10 cm to 60 cm over a century. A much greater change in water levels is experienced daily along most ocean coasts in the form of tides.

Tides and Surges

Davies (1973) showed that about one-third of the world’s shorelines experience 0-m to 2-m tides (microtides), the next third experience 2-m to 4-m tides (mesotides), while the remaining third have tides >4 m (macrotides). Tides themselves do not move significant amounts of sediment but the currents they generate in constricted bodies of water (such as confined inlets, straits, or narrow embayments) can scour the bottom and cause slumping along channel banks. Tides on open coasts control the water level at which waves strike the shore.

Where tides are absent (such as in the U.S. Great Lakes), other factors can affect water levels at the shore. Rainfall and runoff change ground and surface water levels episodically. Winds moving across open bodies of water have the potential to push water up lee shores. These wind tides can be many times higher than astronomic tides at the downwind ends of shallow lagoons such as Laguna Madre (Texas) (Fisk 1959). Winds tilting the surface of a lake can trigger back-and-forth oscillations, called *seiches*, which for a short time resemble the tides along ocean coasts. Lake Erie (U.S.) experiences water-level gradients upward of 4 m between Toledo (Ohio, west end) and Buffalo (New York, east end), when strong west winds blow parallel to the lake’s long axis (Saville 1953, USACE 1995).

The degree to which winds set up water levels at the coast is directly proportional to wind speed and *fetch* (the water distance over which the wind blows). A descriptive term for this is *surge*, technically defined as the “excess water” level above the astronomic tide. Because winds also blow toward the offshore direction, they can produce a negative surge, lowering the water level at the windward shoreline. Highest surges are associated with tropical storms.

Researchers have made great strides in predicting the tide and surge levels associated with storms of particular magnitudes (cf – FEMA 1986). Sophisticated computer models inventory historical storms for a given region, prepare a suite of statistical storms having a certain probability of occurrence over standard periods (such as once in 10 years, 25 years, or 100 years), then determine the effect of each synthetic storm on water levels (tide plus surge) and waves along the coast (Fig 3). These remarkable models are updated and recalibrated as more storm data and historical surge levels are obtained. A practical product of this research is a set of storm-inundation maps that governments use to set minimum building elevations, flood insurance premiums, and related controls on development at the coast.

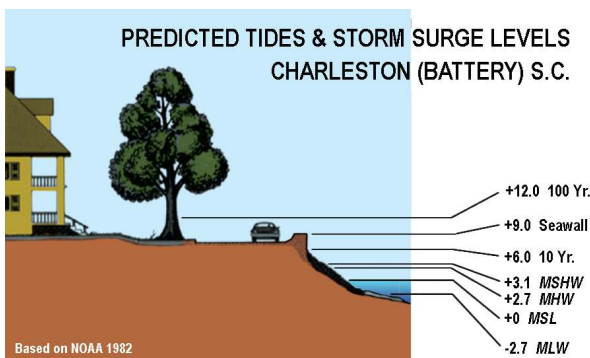


FIGURE 3. Water levels at the coast are influenced by tides, storms, and topography of the offshore zone. Units are feet relative to local mean sea level.

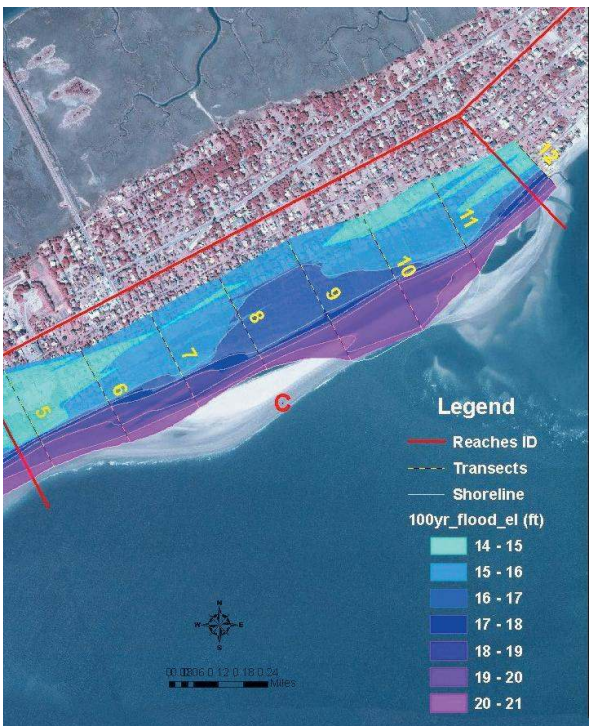


FIGURE 4. Predicted variation in the maximum 100-year flood elevation across a broad oceanfront of a South Carolina barrier island. [Source: CSE/SW/Dewberry 2010]

Storm tide levels are a combination of several components including the astronomic tide, barometric pressure, wind setup, and wave setup. Nearshore bathymetry and the shape of the coast also play a key role with surges “funneling” higher into shallow embayments. As surges strike the land, they will overrun low areas and flood estuaries and lagoons to much higher levels than normal. Where the land is higher, surge elevations will dissipate by varying degrees with distance from the shoreline according to the presence of dunes and the density of vegetation. Even the concentration of buildings and paved areas are accounted for in today’s predictive models. Figure 4 shows the likely dissipation of a 100-year storm surge across the oceanfront of a barrier island in South Carolina. Peak surges of ~20 ft (6 m) at the beach line reduce to ~15 ft (~4.5 m) within 500–1,000 ft (~150–300 m) inland. These results point to the two most important shore-protection strategies for coastal development—elevating structures above the surge level on sound pile foundations, and setting back buildings some distance from the beach.

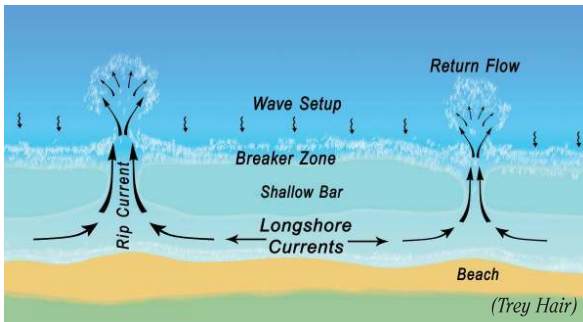
Waves and Wave-Generated Currents

While tides and surges are the principal controls on water levels at the coast, waves do most of the work of moving sediments. Waves arriving at the shore are transformed in shallow water, becoming steeper in the crest and flatter in the trough. As waves approach depths of water similar to their height (measured from crest to trough), they break. The form of the breaker varies from a gradual spilling over at the crest to a gentle up-and-down sloshing. An intermediate breaker type is the familiar plunging (or surfers’) wave. Breaking waves form a bore of water that is propelled toward the shore, running up the slope in proportion to the wave’s height and period (the time for one wave to pass a point). The uprush is followed by a return flow (backrush) toward the next incoming wave. A common name for this process is “undertow.”

Wave-breaking generates oscillating currents (such as the uprush and backrush) as well as circulation currents parallel and perpendicular to the shore. If waves arrive straight to a shoreline, the principal motion is onshore-offshore. However, when waves arrive at an angle to the shore, the motion becomes saw-toothed. Wave-generated currents can be resolved into longshore components as well as cross-shore components. The current associated with the shore-parallel component is simply the *longshore current*. It gains its momentum during the process of breaking so, characteristically, it tends to be strongest between the initial wave breakpoint and the point where the backrush meets the next incoming wave.



Complex “circulation cells” develop in the presence of waves at the shoreline. Water levels tend to “set up” inside the breaker line due to the momentum of waves. This setup is relieved by concentrated seaward flows in the form of “rip currents.” The lengths of rip currents are seldom more than a couple times the width of the breaker zone; hence, the longstanding advice to escape them by swimming parallel to the beach. For more information about wave processes, Basco (1985) and Komar (1998) present excellent descriptions.



Energy Classifications of Coasts

Water levels and waves are such fundamental parameters that broad coastlines are often referred to as tide-dominated, wave-dominated, or mixed-energy—after pioneering work by Hayes (1979) and others. Sedimentary coastlines, where tides are low, tend to be more linear with beaches formed by waves and their alignment matched with incoming wave crests. Tidal inlets tend to be widely spaced, and the principal energy shaping the coast is derived from waves—hence, *wave-dominated coast*. Examples are Padre Island (Texas), the Outer Banks of North Carolina, and Fire Island (New York). In high tide-range settings, waves have less time to work at a particular level. The energy of tides maintains more inlets and breaks up any shore-parallel bars or barrier islands into more shore-perpendicular features. Large tidal volumes entering and exiting lagoons and estuaries each day scour and maintain



channels and generally exert more control on sediment movement than waves—hence, *tide-dominated coast*. Between these two ends of the spectrum are mixed-energy coasts, where waves and tides combine to control the shape of the shoreline. South Carolina is a classic example of a mixed-energy coast (Hayes 1994).

Hayes (1976) coined the term “drumstick” barrier island for the short, stubby barriers of South Carolina (as well as those of the Copper River delta in Alaska and the German Bight in the North Sea). The drumstick shape derives from major accumulations of sand trapped at the upcoast end of the island in the lee of offshore shoals in the adjacent inlet. The downcoast end receives sediments and tends to elongate through the process of spit growth. Before exploring ways the principal energy at the coast controls erosion, a description of common shoreline features is in order.

Littoral Zone

By now, it should be apparent that the shore can actually extend over a broad zone at the coast. At any point in time, the *stillwater level* (ie – the level in the absence of waves) can equal the average global sea level, be well below it at low tide, or be well above it at high tide. It can actually penetrate inland over the coastal plain for brief periods during storms. To understand the basic processes of coastal erosion, one must consider a set of boundaries. At mesoscales, the coastal width of interest is referred to as the *littoral zone* (Fig 5).

This zone is generally defined as the area over which waves in the presence of changing water levels dissipate most of their energy. In common practice, the littoral zone extends from the point of maximum yearly uprush of waves to some small distance seaward of the breakpoint of the largest yearly wave. Along sedimentary coasts, the continual exposure to wave breaking and fluctuating water levels rearranges sediment particles. This leads to development of slopes and morphologic features balanced for the particular waves striking the beach and distribution of sediment grain sizes.

Viewed in cross-section (as in Fig 5), the littoral zone at a site develops a *profile* that is related primarily to sediment texture, wave climate, tide range, sediment supply, and prevailing winds. Key elements of a profile include the following (viewed from a wave’s perspective).

The *outer surf zone* is the gently sloping inshore area over which waves of all sizes begin to break and measurably redistribute sediment. It sometimes includes breakpoint or “longshore bars,” which trigger wave breaking in storms, and troughs between bars. Typical water depths are 1 m to 6 m below sea level. Sediments tend to be finer than the beach but rarely muddy because of the degree of turbulence and mixing that occur in this zone.

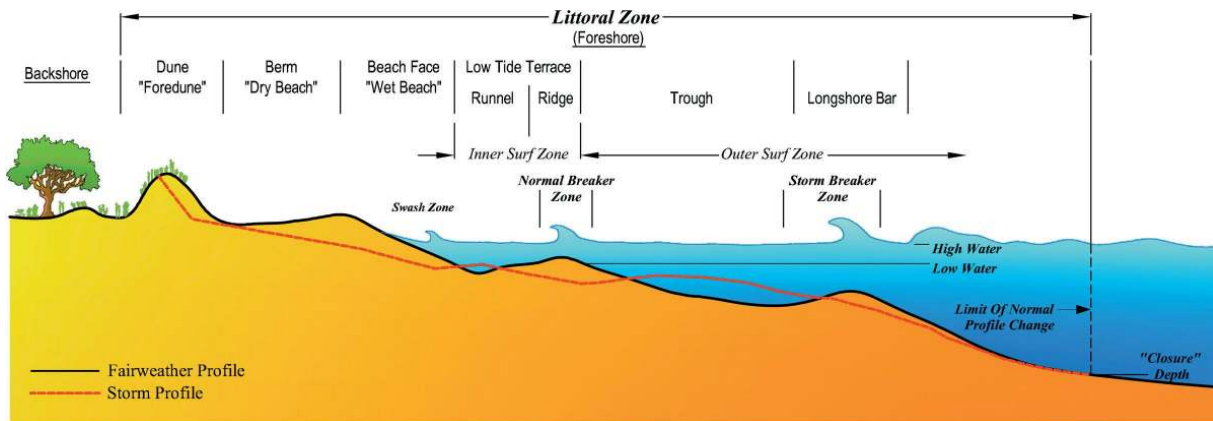


FIGURE 5. Features of the littoral zone. [Source: Komar 1998]

The *inner surf zone* is the area of complex topography between the normal point of wave breaking and usual limit of wave uprush along the beach face, sometimes encompassing an inner bar (“ridge”) and trough (“runnel”) that are exposed at low tide. This zone experiences the greatest vertical change and irregular bottom topography from day to day.

The *beach face* is that portion of the inner surf zone over which wave uprush and backrush occur. It is generally an area of uniform slope that is balanced according to the local sediment grain size and wave climate. This is the final zone of wave energy dissipation and is sometimes referred to as the wet-sand beach over which tides migrate. Coarsest sediments in the littoral zone generally occur at the lower beach face where the plunging action of breaking waves produces the greatest turbulence.

The *berm* is a nearly horizontal portion of the profile beginning at the upper beach face and extending landward to the base of the dune or backshore environment. Situated at the highest wave uprush level, the berm is dry for most of the tidal cycle and therefore is often referred to as the “dry-beach” zone. Typical elevations of the berm are equal to local mean high water plus twice the local mean wave height.

Foredunes are windblown deposits having relatively steep slopes beginning at the landward edge of the berm where finer (noncohesive) sediments accumulate by onshore winds. Terrestrial vegetation establishes itself beginning at elevations that are infrequently flooded, often at a uniform minimum elevation. This offers a distinct demarcation between the “active” littoral zone and adjacent “highland.” Coastal erosion is often measured in relation to movement of the seaward vegetation line because it is a convenient, visible point along the profile. Where foredunes are missing and backshore elevations are similar to those of the berm, *washovers* extend inland some distance, receiving new sheets of sediment with each storm surge. Washovers remove an amount of sediment from the active littoral

zone, forming a base for a possible new dune line along the backshore. Narrow peninsulas (such as barrier islands) may have washovers fanning landward all the way across them to the interior body of water or marsh (Fig 6). Where backshore elevations are much higher than the berm, washovers cannot form. Instead, wave swash cuts away the land, leaving near-vertical escarpments, and transporting new sediment into the littoral zone. Scarps at the edge of the vegetation line mark an abrupt transition between the highland and littoral zone. Vertical scarps in dunes are an indicator of very recent coastal erosion.



FIGURE 6. Washover barrier beach, marsh-filled lagoon, tidal tributaries, and forested beach ridges – common features of the coastal zone. Edingsville Beach, SC, USA

There are many smaller scale features within the littoral zone, including ripples, beach cusps, berm runnels, rip channels, and wrack lines that can change day to day, but are of less importance at mesoscales. Interested readers should consult Davis (1985), Hayes (1994), or Short (1999) for a complete description of beach morphologic features.

Proper assessment of coastal erosion requires study of how the entire littoral zone responds to waves, tides, and currents over a period of time. Most studies through the 20th century evaluated coastal erosion in terms of linear shoreline changes; that is, the displacement of the

shoreline at a single-contour elevation. However, it should be evident that many parts of the littoral zone could be used as reference points. Indeed, historical studies have used the seaward vegetation line, the toe of the foredune, storm debris lines, local mean high water, the berm crest, mean low water, and of course, mean sea level. As long as the comparisons are done rigorously, a reasonable estimate of change is possible. In practice, however, using only one contour can bias the result. This is because the littoral zone is continually adjusting to changes in wave energy.

Measurement of coastal erosion is most problematic using contours on the gently sloping portions of the littoral profile because minor changes in slope lead to large horizontal displacement of the contour. As will be seen in the next section, adjustments in the slope of a beach can produce landward movement of the dune line and seaward movement of the low-tide line—at the same time.

Profiles of Equilibrium

Profiles across the surf zone evolve toward a certain dynamic equilibrium. They are dynamic in the sense that they never completely stabilize under the constant sloshing

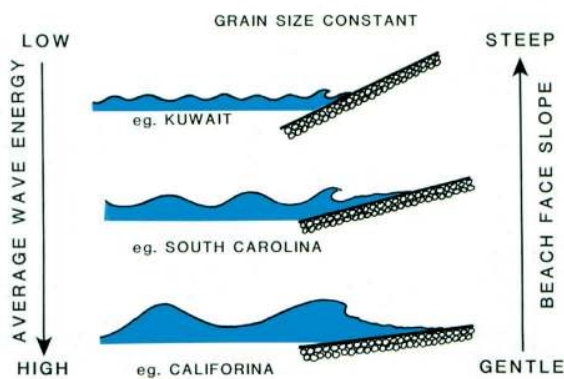
of waves. Yet over days to months, they tend to take on shapes and slopes balanced for the average wave energy and tide levels. Researchers have developed some useful relationships for littoral profiles based on local wave energy and sediment grain size (Bascom 1951), wave steepness (King and Williams 1949), surf processes and slope parameters (Battjes 1974), sediment transport formulae (Dean 1973), and littoral sediment volumes (Kana 1993).

Scientists from Scripps Institution of Oceanography (USA) (eg – Bascom 1951) were among the first investigators to measure systematic variations in beach profiles. Working on plans for amphibious landings during World War II, researchers noted two important relationships: (1) beach slope becomes flatter as wave heights increase and (2) beach slope increases as sediment grain size increases. The first relates to how wave energy is dissipated at the shore. Larger waves have more energy to expend and therefore tend to flatten the profile, so that breaking and energy transfer can occur gradually over a wider surf zone. This is accomplished by a redistribution of some of the sediments from the upper beach face and berm to the lower beach face and inshore zone on any given sedimentary coast. Beach face slope is gentler than normal after storms, for example. As the slope flattens, the character of the breaker changes from plunging to spilling, an important result that reduces the amount of sediment picked up and moved by later waves (Kana 1979).

Changes in beach slope can result from a single storm or occur gradually over several months. Along many coasts, wave climate changes seasonally, giving rise to the terms “summer beach” or “winter beach” (Bascom 1951). Each profile is adjusted for the typical waves occurring in those seasons. Along other coasts, the passage of storms controls the timing of profile changes (Hayes 1967). In either case, the “winter” or “post-storm” profile will tend to be flatter and the dry beach narrower. The summer or “pre-storm” profile will tend to be steeper and the dry beach wider. These systematic slope changes are referred to as the “beach cycle.” They represent a redistribution of littoral sediments in the cross-shore direction in response to varying wave energy.

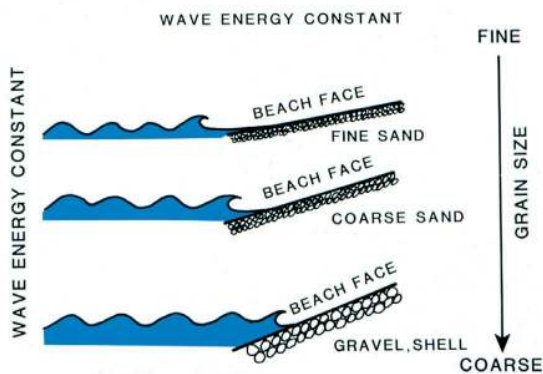
To an observer standing at the base of the foredune, redistribution of sediment from the berm to the low-tide terrace during a storm will look like coastal erosion. And indeed it is, at micro time scales of a few days. But if the berm rebuilds over the next several weeks and is restored to its pre-storm condition, the net result may be no measurable change in the shoreline position. Coastal erosion, remember, is time-dependent. So at mesoscales, it must take into account the range of seasonal or storm profile adjustments around the “equilibrium” or average profile. This is why it is advisable to compare the shoreline condition using data from similar seasons or when the shapes and slopes of the profile are similar.

RELATIONSHIP OF WAVE ENERGY, AND BEACH SLOPE



First Relationship – Beach slope vs. wave energy

RELATIONSHIP OF GRAIN SIZE AND BEACH SLOPE



Second Relationship – Beach slope vs. sediment grain size



Seal Rocks Beach, Australia

The second relationship of importance is beach slope and sediment grain size. Coarse-grained beaches tend to be steeper than fine-grained beaches because of differences in percolation rates in the wave uprush zone. Incoming bores from breaking waves move sediment toward shore, literally building beaches and berms as high as the wave swash can reach. The backrush returning under the force of gravity carries sediment back to the lower beach. In general, the uprush has more sediment-carrying capacity than the backrush because some of the water mass percolates into the sediment pore spaces and is removed from the surf zone before its seaward return. Because coarse-grained beaches are more porous, they absorb wave uprush and accumulate sediments more efficiently, allowing steeper slopes to develop. The limiting factor is gravity. Cobblestone beaches may have equilibrium slopes approaching one on five, whereas very fine sand beaches may have slopes of 1 on 200. Most beaches worldwide have slopes in the range 1 on 10 to 1 on 40.

With these two relationships, it is possible to distinguish whether a particular beach is in equilibrium, or at one extreme or the other of the beach cycle. Early researchers found that profiles are more likely to become “erosional” if wave steepness (wave height/wave length) exceeds a factor of 0.025. Closely spaced, large storm waves meet this criterion. At steepness values less than 0.025, the profile is more likely to build. Other 20th century scientists introduced sediment grain sizes (fall velocities) and beach face slopes to refine these predictors (Dean 1973, Battjes 1974, Komar 1998).

Many coasts are dynamic, experiencing a great variety of waves. Australian researchers and others have shown that beaches and inshore zones move through a number of complex morphological stages related to onshore or offshore accumulation of sediment (Wright and Short, 1984). These changes are most noticeable along moderate- to high-energy beaches composed of a broad mixture of sand sizes. The beach cycle is poorly established along sheltered coasts (such as inland waterways and small lakes) because there is little or no wave energy most of the time.

Considering how complicated and variable the littoral profile can be, determining the true rate of coastal erosion at a site is difficult. Much of the variability, however, can be eliminated by using “profile volumes” to define conditions at a particular shoreline (Kana 1993). Profile volumes are defined as the quantity of sand in a representative cross-section of the littoral zone between the foredune and estimated outer limit of bottom change (Fig 7). By systematically surveying many cross-sections on a given beach, a statistical composite profile can be determined.

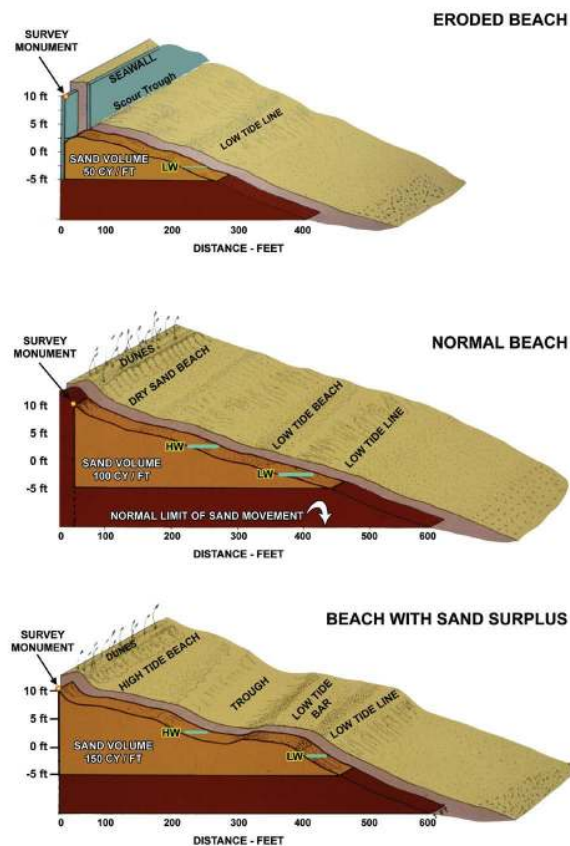


FIGURE 7. Profile volumes—the quantity of sediment within defined boundaries of the littoral zone—provide objective measures of the health of a beach. [Source: Kana 1993]

Its profile volume between defined contours is calculated as the volume of sediment over one-unit distance in the longshore direction.

Profile volumes are independent of the beach cycle because, in effect, they integrate all the microscale features in the littoral zone. As Figure 7 illustrates, profile volumes are arbitrary depending on the reference contours chosen. But their value lies in allowing comparisons from one locality to another. Thus, a profile backed by a seawall (Fig 7, upper) may contain much less sediment per unit length of shoreline than a normal “healthy” profile. Conversely, a profile may contain much more sediment than normal if it incorporates offshore bars. The presumption is that the “normal” profile contains a sufficient volume to allow the normal range of seasonal or storm beach changes without adverse impact to the backshore. In other words, it is in dynamic equilibrium.

Profiles containing only half the normal volume on a given beach (with similar exposure to waves and tides) exhibit a volume deficit and cannot possibly be in equilibrium. In fact, such profiles often have bulkheads or seawalls lining the backshore to prevent further shoreline retreat. If the bulkheads were removed, the shoreline would retreat to the point where there was the minimum profile volume of a healthy beach.

The third type of profile that contains excess sediment volume in nearshore bars is likely to build seaward in the near future. Wave-breaking over these flat, nearshore



Beachgoers crowded against a seawall on a narrow, wet-sand beach

features produces a bore of water that moves shoreward carrying sediment toward the beach. This process continues until the bar “welds” to the beach (Fig 8).

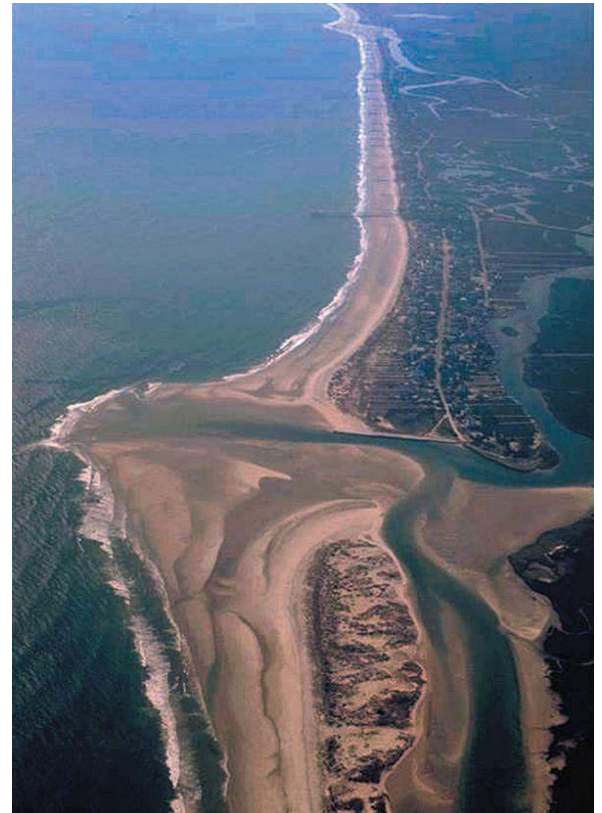


FIGURE 8. Spits (foreground) accrete by sand accumulation and the eventual welding of bars to the shore. Longshore currents carry sand toward the spit and waves push the excess onto the beach. Midway Inlet, SC, USA

In short, coastal erosion should be considered in the context of changes in the littoral profile rather than changes in the position of a single contour. Several relationships exist for evaluating profile development in relation to wave energy, sediment grain size, and other parameters. However, at mesoscales, profile volumes provide an objective direct measure of the littoral condition, effectively integrating most of the microscale variations associated with the beach cycle. If profile volumes at a site decline systematically over time, erosion is occurring, even if the shoreline has been fixed in place by a seawall. If they increase, the shoreline is gaining sediment and moving seaward.

Longshore Distribution of Sediment

So far, the focus has been on the onshore-offshore evolution of the littoral profile and movement of sea level, tides, and waves across this zone. But because coasts have open boundaries, sediment redistribution also takes place in the longshore direction. The primary driving force for longshore transport is breaking waves arriving obliquely at the shore. Not to be confused with coastal currents

that can result from winds (“surface drift”) or major current systems such as the Gulf Stream off south Florida, longshore currents are formed by waves inside the breaker line. Obliquely oriented uprush and backrush combined with excess momentum in breakers add a longshore component to the cross-shore motion in the surf zone. The strength of the longshore current is directly proportional to the breaker’s height and angle of incidence to the beach. High-angled waves generate the fastest longshore currents with peak velocities found just seaward of the mid-surf position (Komar 1998).

Along some coasts or at the extremities of littoral cells, wave direction changes little from season to season, thereby producing one predominant longshore transport direction. Natural indicators of the dominant sediment transport direction include sand spits. Unidirectional transport is particularly common along beaches in low latitudes where easterly trade winds (which generate waves) have relatively constant direction. In mid-latitude coasts such as the U.S. east coast, wind and wave directions are more variable. Westerlies prevail but they are locally modified by the sea-breeze cycle, which intensifies the onshore component as the land heats during the day. The frequent passage of extratropical storms (eg – northeasters) generates opposing waves from the north. Such coasts are therefore subject to alternating periods of winds from opposing directions and longshore transport in either direction.

Longshore transport carries sediment from one section of beach to another. The originating source is referred to as “updrift” and the receiving area is “downdrift.” Therefore, if the shoreline is to remain stable, an equal amount of sediment has to be brought into a section from updrift to balance the loss downdrift. Some have referred to the longshore transport system as a “river of sand” along the coast (Inman and Bagnold 1963). But unlike rivers of water, the direction of flow can reverse with wave direction. Coastal erosion occurs if the amount leaving a segment of the shoreline exceeds the quantity introduced from updrift. The concern at mesoscales is how closely the longshore transport system balances in either direction in response to a full spectrum of wave and tide conditions spanning the seasons. Coasts in equilibrium at mesoscales usually have low, net longshore transport rates (ie – the vector sum of transport moving left and right when viewed from shore approaches zero). Such conditions can occur on low- as well as high-wave-energy coasts. The balance can also be achieved even if gross transport rates (the absolute sum of transport moving left and right) are high.

Equilibrium Planforms

Perusal of coastal maps from around the world reveals great variety in the shape and orientation of shorelines. While some segments such as the south shore of Long Island (NY) appear regular with long, straight barrier beaches, the majority are much more irregular. The



Edisto Beach (SC) – Net longshore transport is from top to bottom and “into the bight” along this barrier spit. Scalloped vegetation line marks locations of groins buried by a recent nourishment project. [Source: CSE]

general planform of a particular coastal segment is strongly influenced by local geology (Davis 1994), but at mesoscales is principally controlled by coastal processes and sediment supply (Hayes 1976, 1994). Something must anchor the segment, otherwise sediments will be dispersed rapidly. Once anchored, the planform will adjust to the incident wave climate, evolving toward an equilibrium where the cross-shore profile changes balance and net longshore transport approaches zero. When mesoscale changes in littoral volumes and shoreline planform approach zero, the segment is said to be in equilibrium.

At mesoscales, shorelines can be anchored by consolidated deposits (such as rock outcrops) or by unconsolidated deposits (such as river and tidal deltas). These features function as “headlands,” creating littoral boundaries and segmenting the coast in between into “cells” (Inman and Bagnold 1963). Loose sediments trapped between headlands are shaped by waves into beaches. Because headlands by definition protrude from the land, they locally alter waves, causing early breaking and “refraction” around the protrusion (Fig 9). The remaining waves between headlands become curvilinear in orientation to a degree related to the separation distance between headlands as well as the size and period of the waves and inshore depth of water. An indication that the shoreline between headlands is close to equilibrium is the angle of wave breaking. If



FIGURE 9. A series of “pocket” beaches bounded by rocky headlands and linked together in a classic “half-moon” bay. The primary headlands flank the entrance to the bay, modifying waves via the processes of refraction and diffraction. The resulting waves fan into the bay with a curvature that matches the beach. Bay of Islands, New Zealand

wave crests parallel the shoreline, breaking occurs at nearly the same instant from one end of the segment to the other, wave angle to the shore is near zero, and there is little or no longshore transport. The equilibrium planform is always concave toward the sea, and the curvature of the beach generally increases with decreasing separation of the headlands. If the incident waves arrive at an angle to the shoreline segment, a “fishhook” planform results. Silvester and Hsu (1994) refer to such shorelines as crenulate-shaped with a log-spiral form that signals the prevailing wave and net drift direction (away from the hook end).



Hawks Nest Beach, Australia (net drift SW to NE)

River deltas and their related forms—ebb-tidal deltas¹—serve as submarine headlands (Fig 10). They produce the same effect on incident waves, causing breaking and refraction around them. But in this case, the unconsolidated sediments of the deltas are also subject to transport. The degree to which adjacent shorelines are in equilibrium, therefore, depends on stability of the delta as well as adjustment of the coast between these dynamic headlands. Many of the world’s great beaches are bounded by unstable deltas. In general, they tend to be more mobile than “equilibrium beaches” bounded by rocky headlands, but they also may receive new supplies of sediment introduced at river mouths.

Limits of the Littoral Zone

Along most sedimentary coasts, nearly all measurable change in bottom topography at decade to century time scales occurs in relatively shallow water. This is an important finding because accurate measurements and prediction of coastal erosion depend on a defined outer limit of the littoral zone. Early studies in the 20th century noted the ability of large ocean waves to produce orbital motions sufficient to move sediment in deep water. Wave

¹Footnote: The term ebb-tidal delta (Hayes 1975) refers to the seaward shoals of an inlet, which are deposited and controlled by the outgoing or “ebb” tide. The counterpart is a flood-tidal delta deposited in the lagoon by the flood tide.



FIGURE 10. River and tidal deltas act as mobile, submerged headlands, anchoring littoral cells in between. Captain Sams Inlet, SC, USA

motion in depths >30 m can be sensed by divers and observed to form ripples on the bottom. However, the sloshing motion of large waves in deep water is quite symmetrical and therefore unlikely to result in significant net movement in one direction. Sediment transport in deep water tends to move little measurable volume which can be detected as a change in bottom elevation—at least when compared with the elevation changes in the littoral zone. The sediment transport of importance to coastal erosion mostly occurs in shallow water close to shore. Not surprisingly, considerable research effort has been applied to define this zone.

The normal seaward limit of measurable bottom change in a beach profile along the coast is defined as the *depth of closure* (DOC)—the term derived from the observation that a set of comparative profiles will tend to converge some distance offshore. Like coastal erosion, DOC is time-dependent and site-specific. It will tend to be deeper as more time elapses, and the effect of more storms or higher waves applies. DOC can be estimated several ways at a site—using empirical formulae (Hallermeier 1978), measuring sequential profiles, comparing historical charts, or marking sand/mud transitions offshore.

Figure 11 illustrates the use of sequential surveys to estimate DOC for a site on the south shore of Long Island (NY). A series of littoral profiles were surveyed at one station between 1979 and 2003. Each profile was overlain and compared. There is considerable variation in elevation across the active beach and outer bar, indicating large changes in the bottom. However, beyond a depth of ~6.2 m (~20 ft) below approximate mean sea level (the datum used in the surveys), the bottom elevation remains relatively constant with changes in the order of 10 centimeters from survey to survey. Such small changes are in the range of normal survey error for quality profiles. Thus, for this particular site over a 24-year period, it is accurate to say that nearly all measurable bottom change occurred between the dune line and the ~6-m depth contour. Surveys extending to at least that depth would tend to

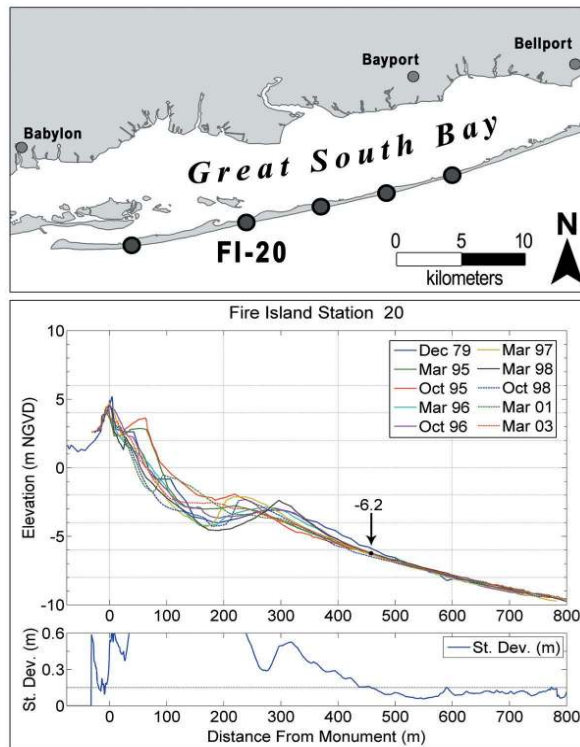


FIGURE 11. A 24-year series of profiles at Fire Island, NY, USA, overlain to illustrate the concept of depth of closure (DOC). The lower graph shows the standard deviation in elevation reduced to <0.15 m (6 in) at a point 450 m offshore. This yields a DOC of ~6.2 m (~20.3 ft). [Source data: NY Dept of State and USACE]

account for most of the erosion or accretion at that point along the shoreline.

DOC will be shallower in lower wave-energy settings. A segment of the South Carolina shoreline, illustrated in Figure 12, is characteristic of a mixed-energy setting with lower wave heights (and higher tide range) than the Long Island coast. The red zone on the figure inside the -6-ft mean lower low water (MLLW) contour is where ~90 percent of the active littoral sand transport occurs. The orange zone to -12-ft MLLW is where most of the remaining sand exchange occurs between the beach and the inshore area (Fig 13). The -6-ft and -12-ft MLLW contours are commonly illustrated on nautical charts. Given South Carolina's tide range, they equate to ~9 ft to 15 ft (~2.7–4.6 m) below mean sea level. DOC estimates for nearly all US Atlantic Coast and Gulf Coast beach sites are rarely >9 m; hence, the accepted standard profile limit of ~30 ft practiced by the US Army Corps of Engineers (CERC 1984).

The point of this technical discussion will become apparent later in this primer. But it is important to note here that a full understanding of coastal erosion is only possible if the limits of the littoral zone are known at a site. In South Carolina and many other sites, that means close tracking of sediments in the red zone (Fig 12) as well as on the visible beach. And as Figure 12 implies, tidal deltas

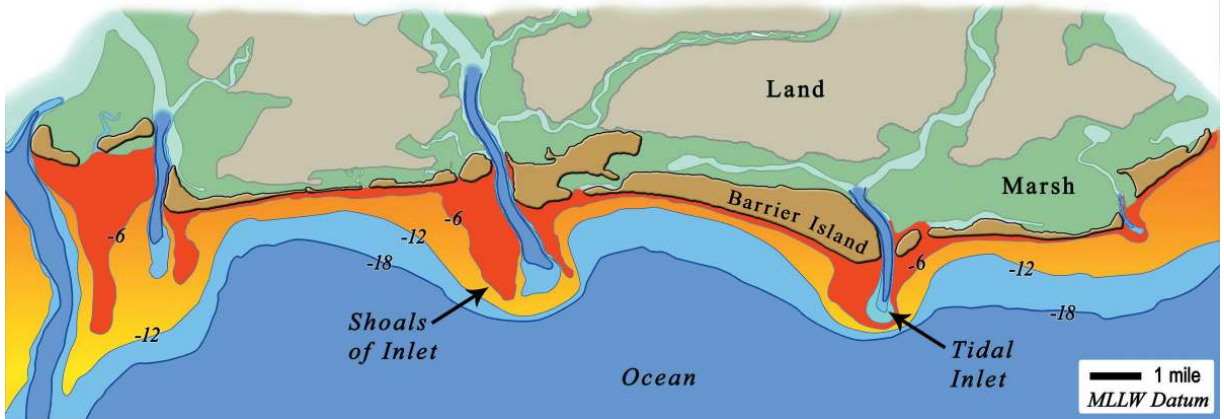


FIGURE 12. A section of mixed-energy coast (Hayes 1994) from central South Carolina showing the principal sand bodies (red and orange areas) of the littoral zone at mesoscales. Depth contours are in feet above mean lower low water datum. Tidal deltas in mixed energy settings like this tend to contain as much volume as the adjacent barrier islands. (Graphic by Trey Hair)

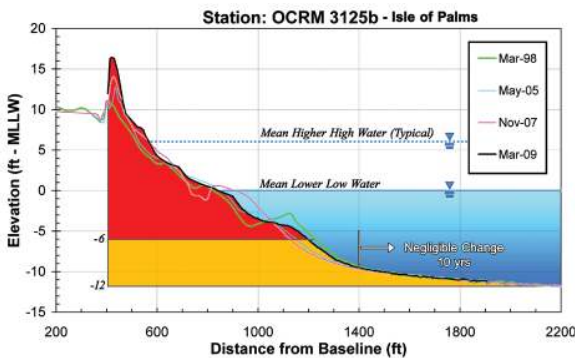


FIGURE 13. Comparative profiles for a central South Carolina beach showing the principal zone of sand movement over an 11-year period in water depths shallower than -12 ft MLLW (~4.6 m below local MSL). [Source data: Courtesy SCDHEC-OCRM]

often contain as much littoral sand as the adjacent barrier islands (Hayes 1994).

Shoreline Salients

While it is generally true that equilibrium shorelines will be concave in planform in the seaward direction, locally, shorelines may bulge seaward. Usually, this is the result of a change in the bottom offshore. These shoreline “salients” (Dally and Pope 1986) form in the lee of a bathymetric high, which may be associated with a reef or longshore bar segment, an artificial obstruction such as a breakwater, or the shoals of a tidal delta. Offshore islands produce salients in their lee because they block incoming waves and cause bending around the landmass. Longshore transport becomes interrupted behind the island, leaving deposits that form the salient. At the extreme, the salient may accumulate enough sediment to migrate seaward all the way to the island. The isthmus formed in this manner is called a *tombolo* (Fig 14). There is an unlimited range of scales for coastal salients. Many are dynamic because the offshore feature is mobile. In recent years, spoil from dredging operations in harbors has been placed close to shore and formed into “underwater berms.” Although

intended to feed the beach by migration of dredged sediments up the littoral profile, the berms sometime have a more immediate effect by changing the longshore transport rates and producing salients on the lee shore (Douglass 1997).

Other salients can be produced by artificially interrupting the longshore transport system by *groins* (structures built perpendicular to the shoreline across all or part of the littoral zone), detached breakwaters, or even new inlets (Kana 1989, ASCE 1994). Ebb-tidal deltas, such as the red and orange bulges in Figure 12, produce dramatic variations in waves arriving at the shore. This, of course, leads to more variability in shoreline erosion trends near inlets. Tidal deltas, in many respects, act like groins or jetties—trapping and retaining sand, and holding a segment of shoreline in place. In a mixed-energy setting like South Carolina, the effect of these seaward shoals generally dwarfs the effect of any manmade structures.

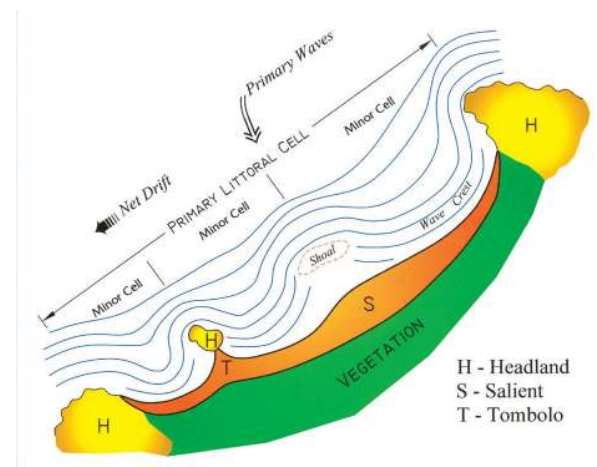


FIGURE 14. Complex coastal land forms are shaped by variations in wave direction and offshore topography. Local erosion is usually linked to regional features and the predominant coastal processes within a littoral cell. [Source: Coastal Science & Engineering]



The first beach nourishment project in the U.S. was along 7 km of Coney Island (NY) in 1923. Image is circa 1940. [Source: Life Magazine]

Coastal erosion at mesoscales is so highly variable, largely because of the complexity of coastal processes (winds, waves, currents, and tides) and complication of offshore bathymetry. To the extent that these are understood at a particular site, it is possible to project future shoreline positions and begin to quantify rates of change at decadal to century time scales.

During the past couple of decades, there have been extraordinary advances in process-based modeling of shoreline change in the presence of offshore bars, shore-protection structures, and tidal inlets. Computer simulations of morphologic change is one of the most active areas of research, building on pioneering work by U.S., European, and Japanese researchers. Some of the practical models include GENESIS (Hanson and Kraus 1989), SBEACH (Larson and Kraus 1989), IRM (Inlet Reservoir Model) (Kraus 2002), and the world-leading Delft3D (Delft Hydraulics Lab-Netherlands). The recent model GenCade (Kraus et al 2010) provides regional simulations of beach/inlet interactions at mesoscales. While computer simulations of coastal processes and shoreline change are highly advanced, there is always need for accurate field data at a site and continued improvements in models.

QUANTIFYING COASTAL EROSION

Coastal erosion is most often reported as a linear change in shoreline position. It may be measured a number of ways but generally is marked in reference to a fixed object on land such as a surveyor's benchmark. A set of points is usually established along the adjacent highland, and the distance to a particular shoreline contour or morphologic feature is measured periodically from each point. This can also be done with less accuracy but broader coverage using

remote-sensed data such as aerial photographs, satellite imagery, or most recently – a sophisticated down-looking radar system called LIDAR (light detection and ranging) (Lillycrop and Banic 1992). Differential global positioning systems (DGPS) are also facilitating shoreline mapping and providing accurate elevation data as well as shoreline position (List and Farris 1999).

All of the linear shoreline change methods make the assumption that the littoral profile will maintain the same shape and average slope over time. As described in the previous section, this is not always the case.

Because coastal erosion is the loss of some finite volume of sediment, it is appropriate to quantify it in volumetric terms. This can be accomplished by obtaining repetitive profiles across the entire littoral zone. While more expensive than linear shoreline change measurements, accurately surveyed profiles provide more representative data. By overlaying successive surveys measured from common benchmarks, the difference in cross-sectional area can be computed. This two-dimensional cutaway of the profile becomes a three-dimensional volume by extrapolating over one unit length of shoreline (cf – Fig 7). Thus, a cross-section difference of 1 square meter (m^2) between two overlain profiles from the same site equates to a unit-width volume change of 1 cubic meter per meter (m^3/m); that is, 1 m^3 is lost or gained over a representative 1-m length of beach. Using English units, an equivalent would be 0.4 cy lost or gained over a representative 1-ft length of beach (ie – 0.4 cy/ft).

Standard practice is to collect profiles every 30–600 m (~100–2,000 ft) along the coast in the area of interest and assume the unit-volume differences from profile to profile vary linearly. Clearly, more accuracy is possible with closely spaced transects, particularly where beaches exhibit rhythmic topography. Overall volume change is estimated by averaging the results from adjacent profiles and applying the average over the intervening distance.

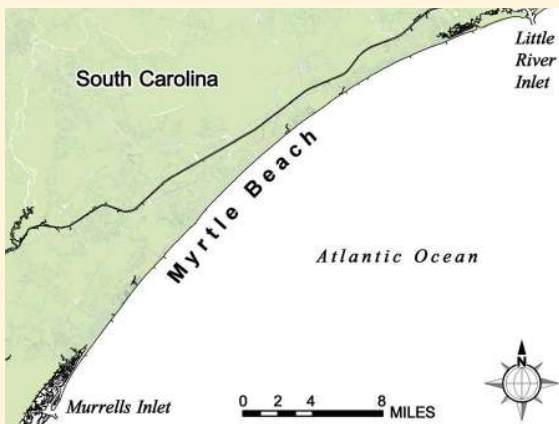
[NOTE: Even with advancements such as multi-beam echosounding, which can provide blanket coverage of the bottom, littoral surveys in relatively shallow water will always be constrained by the time and costs to obtain data. Many more survey lines are required the shallower the area, regardless of technology advances. The energetic conditions of the surf zone or tidal deltas further complicate accurate data collection.]

Summing results for the shoreline reach in question yields a net volume lost or gained during the period between the surveys. A useful practice is to annualize the result and report average unit-width volume change rates (as in $m^3/m/yr$).

Myrtle Beach

Application of Beach Monitoring Surveys

Myrtle Beach, South Carolina, illustrates the utility and importance of annual beach surveys. Situated in the center of a 40-km (25-mile), arcuate shoreline between Little River Inlet and Murrells Inlet, Myrtle Beach is the heart of a booming tourist area. Cottages which lined the shore a century ago have given way to high-rise hotels, amusement parks, and full-service resorts. Visitors look out from their balconies across landscaped patios and pool decks to a naturalized strip of dunes fronting a wide sandy beach. There are no shore-protection structures to be seen - just a continuous expanse of beach extending for miles in either direction. Vistas were not always as attractive though. As recently as 1985, much of Myrtle Beach was armored by seawalls and there was little or no high-tide beach.



There was a perception in the early 1980s that erosion was bad along Myrtle Beach. Yet on closer examination, it became apparent that the problem was not as much due to encroachment of the sea, but rather construction of buildings, pools, and parking lots closer to the beach. To protect their investments, most owners built seawalls, the shore-protection measure of choice in the 1970s. Often, this was the response to minor storms which temporarily cut back any remaining dune along the property. Once in place, seawalls protected upland development but did nothing to preserve the recreational beach.



3rd Avenue South (1985)

Like many eroding beach communities in the U.S., the City of Myrtle Beach requested (in 1977) federal assistance to restore and maintain the beach. But in addition, the City committed to beach monitoring and an interim restoration plan, recognizing that federal nourishment projects

usually involve 20 years of planning and review before implementation. Locally funded studies by CSE demonstrated that Myrtle Beach lost sand between the 1950s and 1980s at low rates of 0.5-2.5 cy/ft/yr (~1-6 m³/m/yr). It was possible to quantify such decadal losses because the U.S. Army Corps of Engineers had obtained profiles along Myrtle Beach after Hurricane Hazel in 1954. CSE was able to locate and duplicate the Corps surveys in the 1980s. These comparative profiles were important because they confirmed low volumetric erosion rates for the area. This finding became the basis for an interim beach nourishment project funded by a local 2 percent accommodations tax. The City wanted to maintain a minimal recreational beach until such time as the federal project received its appropriation for construction.

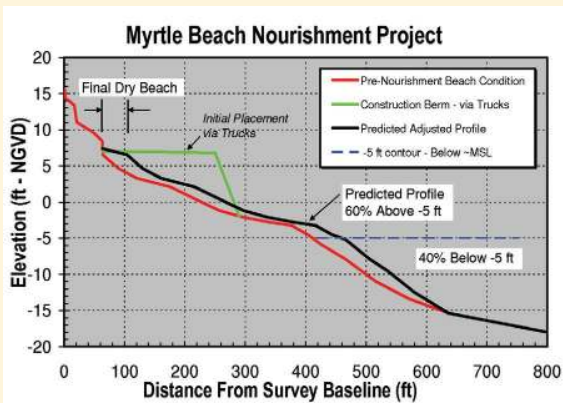


17th Avenue South (1983)

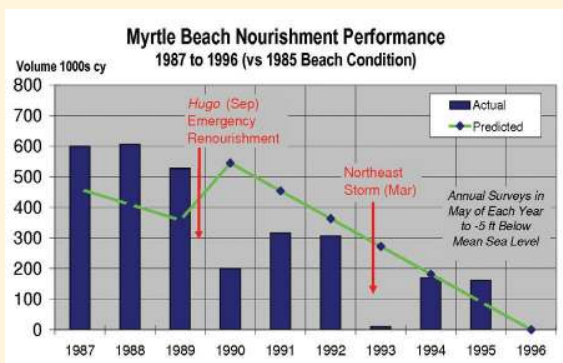
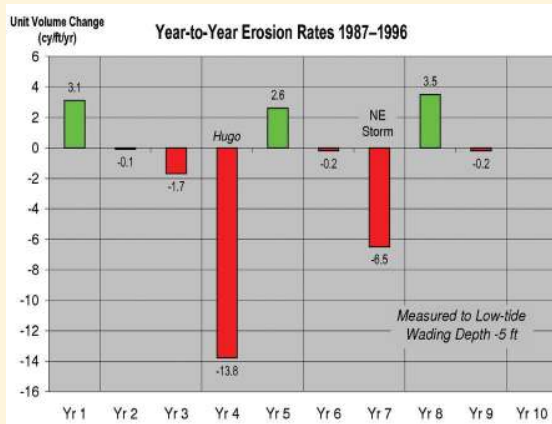
At the direction of the City, CSE prepared a ten-year plan whereby 850,000 cy (~650,000 m³) of beach-quality sand were hauled by truck from inland borrow areas, and then dumped and spread along the beach. Cost of the project was (~)\$4.5 million (1986). While 60,000 truckloads made this one of the largest nourishment projects of its type, the average amount of sand placed was relatively small at ~20 cy/ft (~50 m³/m). The initial sand placement above the low-tide mark was expected to adjust rapidly with nearly half the volume shifting seaward of the low-tide line.

CSE measured the nourishment volume remaining on the beach each year for the next decade using a network of 60 profile lines, many of which extended to 15-ft depths (the estimated DOC). The project was tested by Hurricane *Hugo* in 1989, which removed over half the nourishment sand (measured to low-tide wading depth). Emergency renourishment (396,000 cy via trucks) was performed in 1990 using a FEMA Category G community assistance grant (approximately \$2.5 million) – the type of disaster funding made available to beach towns that have engineered and maintained their own nourishment projects (www.fema.gov/government/grant/pa/re_categories.shtml).

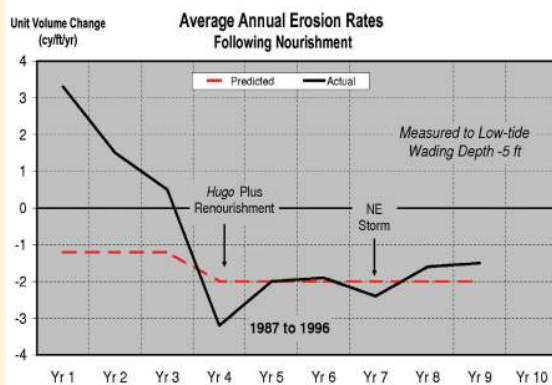




Predicted profile adjustment



Nourished volume remaining vs. predicted



As the accompanying graphs illustrate, the volume of sand in the ~9-mile project area, measured from the seawall line to low-tide wading depth (ie – the portion of beach seen and used by visitors), fluctuated and diminished over time. A decade after the initial project, only about 25 percent of the nourishment sand remained on the visible beach (Kana et al 1997). Yet the average rate of sand loss closely matched the predicted erosion rate over time and was similar to the pre-project historical rates. While year-to-year erosion rates varied considerably, the average annual loss rate steadied within a few years. Much of the variability, in this case, reflects onshore-offshore transport into and out of the measurement zone to wading depth. Surveys to deeper water tend to show less variability and better account for true sand losses.

The City of Myrtle Beach knows how much sand was added to their beach; the community knows how much remains. And city and federal officials know what the rate of sand loss is so they can strategically plan the next project.

When beach erosion is quantified by careful surveys over a sustained period, it becomes possible to project the volumes and costs of restoring, maintaining, and advancing the shoreline. Places like Myrtle Beach have two advantages over many beach communities: their erosion rates are low, and the value of oceanfront property is high. On a per foot per year basis, the cost of nourishment in this case is dwarfed by the value of properties at risk.

In round numbers, the City of Myrtle Beach and FEMA spent (~)\$7 million (1990) to place 1,250,000 cy along ~9 miles of shoreline to keep pace with erosion between 1986 and 1996 until a federal project could be implemented. This equates to <3 cy/ft/yr added at a cost of less than \$20/ft/yr. In 1997, the long-awaited federal nourishment project placed ~2.2 million cubic yards at a cost of (~)\$17 million. And in 2008, as planned, another 1.5 million cubic yards were added to Myrtle Beach to advance the shoreline even further seaward.



10th Avenue South (2009)

The City of Myrtle Beach continued to monitor the beach every year by way of annual profile surveys across the littoral zone (CSE 2010). These data confirm that the City has received ~4.4 million cubic yards in four nourishment events (1986, 1990, 1997, and 2008), and there are now ~2.5 million cubic yards more sand on the visible beach than in 1985. On average, the beach is 80-100 ft wider today than it was 25 years ago. Nearly 75 percent of the placed sand remains in the project area out to the DOC (estimated to be -15 ft MSL in this setting). Average annual erosion rates of the nourishment sand have been ~1.5 cy/ft/yr since the 1980s.

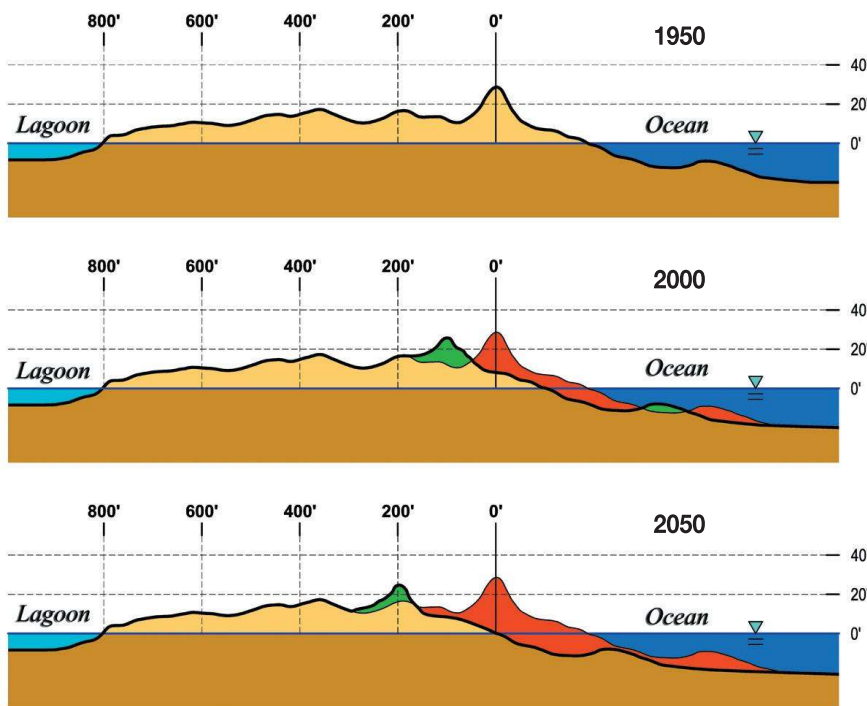


FIGURE 15. Cross-sections through a barrier island eroding at an average rate of 2 ft/yr. Note preservation of littoral zone morphology. Such recession equates to a 50-year volumetric loss of ~100 cy/ft. (See text for additional discussion.)

The advantage of volumetric analysis for mesoscale erosion studies is that it provides a direct estimate of sediment quantities. Linear shoreline changes may be determined using the same survey data, simply by selecting common contours and comparing their displacement between surveys. With sufficient site-specific data in hand, it is possible to compute simple statistical ratios between volumetric changes and linear shoreline changes. In the early 20th century, U.S. engineers used a simple rule of thumb for estimating volume loss or nourishment requirements as a function of beach area (width) (CERC 1984). They found for Coney Island (NY) that a one-foot shoreline retreat equated to approximately one cubic yard volume loss per one linear foot of shoreline. Thus to replace one square foot of beach area, the engineers estimated one cubic yard would be needed.

Measuring coastal erosion in volumetric terms rather than linear terms immediately provides estimates of sediment losses (or beach nourishment requirements to restore previous losses). Profile volumes (as described in a previous section) allow estimates of the deficit or surplus volume on a given section of coast in comparison to a “healthy” (or desirable) profile. Figure 15 shows how such surveys can be applied. The three profiles in the figure are typical cutaways through a barrier island. Using round numbers, assume that the long-term linear erosion rate has been 2 ft (0.6 m) per year. Since ~50 years ago, then, the shoreline has receded ~100 ft (30 m). In the next 50 years, it is expected to recede another 100 ft, leaving it 200

ft (60 m) landward of its initial position (in this time frame). This, of course, assumes erosion rates remain constant.

The sequential profiles illustrate how the morphology of the littoral zone is preserved. At 2 feet per year (ft/yr), a realistic rate in the 20th century for many developed beaches (Dolan et al 1990), dunes have time to reform and migrate inland. The offshore bar and active beach similarly migrate with the profile. The red cross-section, therefore, is a measure of the lost sediment over time. Using the simple rule of thumb that 1 ft (0.3 m) of linear retreat equates to 1 cubic yard per foot (cy/ft) of volumetric erosion (equivalent metric value is 2.5 m³/m), the 50-year loss will be ~100 cy/ft (~250 m³/m), and the 100-

year loss will be ~200 cy/ft (~500 m³/m). **The essence of the beach restoration debate is the question of whether it is better to allow erosion to proceed (and move buildings out of the hazard area) or to replace the lost sand in the red zone (Fig 15).**

Relatively few shorelines worldwide are monitored for profile changes. Yet, along developed coasts, this should be a high priority, considering the property values at risk. Present practice in some U.S. states (including California, Florida, and South Carolina) is to survey networks of profiles along developed ocean shorelines at least once per year (USACE 1995). These types of data sets become more valuable over time because they define the condition across the entire littoral zone and provide more objective information for planning. The accompanying sidebar illustrates how profile surveys were used to plan a nourishment project at Myrtle Beach (SC) and to track its performance.

SIGNATURES OF COASTAL EROSION

The previous section on quantifying erosion, as well as the case study at Myrtle Beach, is simplistic yet applicable to a large number of developed beaches. There are numerous locales where erosion rates are moderately low and fairly uniform over long distances. Most of Bogue Banks (NC), a 25-mile-long (40 km) barrier island, has 50-year erosion rates averaging 2-3 ft/yr (source: NCDENR 1998, 2004). This implies the processes molding and shaping Bogue

Banks are essentially the same over the length of the island. It is only near the inlets that Bogue Banks exhibits a wider range of shoreline changes. Fortunately, most of the developed coast is changing within a narrow range of the order ± 1 m/yr (Dolan et al 1990).

Some researchers have suggested that sea-level rise is the principal underlying cause of coastal erosion (eg – Leatherman et al 1999). Yet only a small fraction of the observed erosion at most sites in the 20th century can be attributed to sea-level rise. For example, a 3-mm/yr rise in sea level (a typical local trend taking into account eustatic plus local effects such as subsidence), applied over a beach profile slope averaging 1 on 30, accounts for about 0.1 m/yr of apparent recession. Obviously, this value is dwarfed by the remainder of the rate in areas where erosion exceeds 1 m/yr. Clearly, other erosion factors must be more important than sea-level rise in most settings—at mesoscales.

Studies from many parts of the world demonstrate that certain erosion-causing factors other than global sea-level rise are the primary controls on a shoreline at mesoscales. Recurring trends and cycles of erosion and accretion have been observed along particular coasts that help identify the root causes of erosion. These “signatures of erosion” (Kana 1995) place a shoreline in context and can be used to distinguish between cyclical, temporary problems, and chronic long-term trends. Following are some examples.

Erosion Caused by Artificial Structures

Some scientists attribute most coastal erosion problems to artificial structures. To the extent that such “erosion control devices” as jetties, groins, seawalls, and breakwaters alter or interrupt littoral sediment transport, they will produce localized effects including accelerated erosion downcoast (Fig 16). The degree of influence of coastal structures is generally a function of scale and how much of the littoral profile is directly exposed to them. Long jetties extending thousands of meters into the sea will have more effect on longshore transport than short groins perched solely across the dry beach. The former may trap and retain millions of cubic meters of sand on the upcoast side (as in the case of jetties for the port of Lome, Togo, on the West African coast) to the long-term detriment of downcoast areas. Short groins, by comparison, will have little effect on waves and currents in the surf zone, leaving sand to move freely along the coast most of the time. These shore-perpendicular structures will often affect adjacent shorelines (positively in the upcoast direction and negatively in the downcoast direction) for distances of 10–20 times their length (ASCE 1994).

Seawalls, revetments, and bulkheads are built along many coasts for shore protection. Like short groins, these shore-parallel structures have little effect on day-to-day coastal processes until they are exposed to waves and littoral



Cape Hatteras (NC) groin – trapping sand moving from left to right (N to S).

currents. Seawalls are often built in response to erosion following a storm event. Not surprisingly, they may be covered up or removed from the active littoral zone if the dry beach recovers after the storm. Along stable coasts, seawalls may remain buried most of the time only to be exposed during the next catastrophic storm many years later. In this sense, they provide silent protection with little impact to the beach. Shore-parallel structures exacerbate erosion, however, once they are regularly exposed to the surf zone. They retain sediment that otherwise would feed littoral flows, and generate reflected waves that



FIGURE 16. Artificial structures (such as seawalls, groins, and jetties) are used to anchor parts of the coast, often with negative as well as positive effects. In some settings, their impacts are dwarfed by natural features like the inlet bar (in the foreground), which is acting as a natural breakwater. Sullivan's Island, SC, USA



Isolated shore protection structures (such as this bulkhead along the New York coast) locally modify erosion rates but do not alter the regional erosion trend. [Photo by M Kana]

induce scour at the toe of the structure. Where bulkheads terminate abruptly, the adjacent, unarmored shore sustains localized erosion from waves diffracting away from the structure's exposed end.

Countless examples exist of erosion caused or accelerated by shore-protection structures. In fact, in some jurisdictions hard erosion control structures are now prohibited, or must be mitigated by means of nourishment (artificial addition of sediment to the littoral system) or mechanical "bypassing" of sediment from the upcoast to the downcoast side of the structure. But if many of today's shore protection structures were built in response to erosion, are they the primary cause of it? Usually, there is an underlying cause of erosion that is fundamental at any particular site and is likely to be operating on a grander scale (Basco et al 1997).

Headlands, Rivers, and Deltas

Because coastal erosion at mesoscales is intimately linked to sediment supply, major causes can be found associated with changes in headlands, rivers, and deltas. Headlands of unconsolidated sediments [such as the glacial till of Montauk (NY) bluffs] provide a prime source of sediment to the south shore of Long Island. As long as sources such as this are available to feed downcoast areas, erosion will be lessened. But when such sources are depleted, erosion is likely to accelerate. Such has been the case with dams on rivers. Many of California's beaches sustained erosion throughout the 20th century because of a reduction in sand supplied by rivers. The California Coastal Commission is presently investigating ways to restore at least portions of the supply by removing dams or dredging sediments that are clogging upstream reservoirs.

Unquestionably, the largest supplies of sediment at the coast are found at river mouths. Deposition is more often associated with deltas, particularly the Amazon and Mississippi Rivers, given their enormous discharges. However, unconfined rivers across the coastal plain have a tendency to change course, distributing their flows

among several channels. If flows wax and wane among several distributary channels, the sediment supply to the littoral zone will also fluctuate. Bars and barrier islands in one portion of the delta may receive an influx of sediment for a time, then sustain periods with no new sediment. *Channel avulsion*, or the sudden shift from one distributary channel to another, often signals a switch from deposition to erosion along a particular shore. This (and rapid subsidence) account for high rates of erosion along Louisiana's southwest coast. The path of the Mississippi River switched to its present, more easterly, course more than two centuries ago and no longer feeds the eroding area with new sediment.



Tidal deltas (such as this one in South Carolina) contain large reservoirs of sand, sometimes dwarfing the amounts found on adjacent barrier beaches.

Tidal deltas are also associated with rivers and inlets at the coast. Many inlets persist despite the lack of major rivers discharging nearby. They flush the sounds and lagoons that separate barrier islands from the mainland. Tidal flows maintain channels and lead to formation of deltas on the landward as well as seaward side of inlets. The size of tidal deltas can be impressive, with some containing more sand in their shoals than is found on the adjacent barrier islands. For example, Stono Inlet near Charleston (SC) contains almost 100 million m³ in its ebb-tidal delta, which is enough sediment to build a 25-km barrier island over 1 km wide (Kana 2002). Some studies suggest that much of the erosion or accretion occurring today relates simply to recycling of sediments between tidal inlets and the beach (Hayes 1979). This seems to be the case for Virginia's barrier island shoreline, the west coast of Florida, and other mixed-energy settings.

Navigable inlets in recent years have become flash points for erosion controversies along nearby beaches. On the one hand, safe entrance channels are needed for commerce and trade (as well as recreational boating). But on the other hand, maintenance of channels by dredging or stabilization with jetties often leads to changes in the sediment supply to adjacent beaches. Exacerbating the problem is the common practice of disposing dredged sediments offshore and removing them from the littoral zone. Best management practice now encourages disposal of quality sediments from navigation channels on the adjacent beaches. However, the question remains whether a particular human work is more

responsible for erosion than the natural processes acting on the site. It is not always obvious, particularly when the changes along the shore are relatively small over a period of time.

Other Erosion Signatures at Mesoscales

After the major erosion-causing factors are identified for a site (sea-level rise and subsidence; changes in sediment sources in headlands, rivers, or deltas; or sediment trapping by inlets and coastal structures), there are other common signatures of change that help place the problem in context. Following are some recurring examples.

One-Way Beach Cycle – A one-way beach cycle (Kana 1995) is a shoreline change signature characterized by high erosion rates and related to a gross imbalance between storm wave energy and normal wave energy. This is common along sheltered shorelines such as estuaries and lagoons where fetches limit wave heights (Nordstrom 1992). The highest waves in storms combined with tidal surges cut back the adjacent highland, leaving escarpments and shifting sediment down the littoral profile. After storms, there is insufficient energy to rebuild the upper beach. The net result is a “one-way beach cycle,” with chronic, rapid erosion linked to the frequency and intensity of storms.

Shoal-Bypass Cycle – Sediment *bypassing* is a process whereby sand crosses an inlet or entrance channel and shifts from one beach to another (Bruun and Gerritsen 1958). Once considered a quasi-steady phenomenon, bypassing in many settings is actually episodic. Long periods occur with little or no exchange of sand between inlets and beaches. Then an event such as a storm or channel avulsion may initiate the transfer. Entire sand bars (or shoals) may move onshore and accrete along the downcoast beach—hence, the term “shoal bypassing” (Sexton and Hayes 1983, Gaudio and Kana 2001).

River and inlet deltas contain shoals that coalesce offshore as sediment accumulates. A change in channel orientation may free shoals at the margins of the delta and allow waves to push them onshore. As this process develops (Stage 1 – Fig 17), the shoal acts as a breakwater, allowing a salient to form along the lee shoreline (Kana et al 1999).

The salient initially derives its sediment from the adjacent beach, causing rapid localized erosion. At Stage 2 of the process, the shoal attaches to the beach, adding a measurable quantity of new sediment. Once attached, the bulge produced by the new deposit becomes a focus of wave energy. It is then spread in either direction by longshore transport (Stage 3) until the shoreline straightens and wave angles diminish. During this process, localized erosion can be exceedingly rapid and destroy property not set sufficiently landward. Yet, ironically, the reach in question may benefit over the long run because of the addition of a new supply of sediment.



Murrells Inlet (SC)

The Special Case of Tidal Inlets

Tidal inlets in many settings are the principal erosion factor along adjacent beaches. Inlets and their associated tidal deltas may trap and withhold sediment or periodically release it in the form of accreting bars. Inlet channels impinge on the littoral profile. The degree to which sand “bypasses” across an inlet from beach to beach largely signals future erosion trends.

Inlets may be positionally stable (particularly if they are deep and anchored in older consolidated sediments), migratory (wherein the channel shifts downcoast under the influence of spit growth on the updrift side), or stabilized (by coastal structures, such as harbor entrance jetties – FitzGerald 1996). Breach inlets are created by storms, draw off large volumes, and otherwise destabilize the beach.



Storm breach – Hatteras Island (NC) – 2003

Tidal currents in inlets lead to sediment accumulation at the seaward end (ebb-tidal delta) and the lagoon side (flood-tidal delta). The size of inlets and their associated deltas is proportional to the volume of water entering and exiting each tidal cycle (O'Brien 1969).

Flood-tidal deltas tend to be larger where there are open lagoons (with area to receive sediments) and high wave energy on the open coast which pushes sediments shoreward inhibiting expansion of seaward shoals. Ebb-tidal deltas tend to be larger than (active) flood-tidal deltas where the lagoon is marsh-filled and there is little extra storage capacity for new sediment. This latter situation produces asymmetry of the tides with the ebb flows being stronger than the flood tide. The imbalance creates a net sediment transport in the ebb direction (FitzGerald 1996), which is especially important in mixed-energy settings like South Carolina. Besides allowing ebb-tidal deltas to extend far offshore (see Fig 12), ebb dominance preserves littoral sediments seaward of the coastal strandline (Kana et al 1999). To the extent sand contained in the littoral zone and ebb-tidal delta is not lost to washovers, breach inlets, or flood-tidal deltas, the mesoscale sand volume loss will be lower.

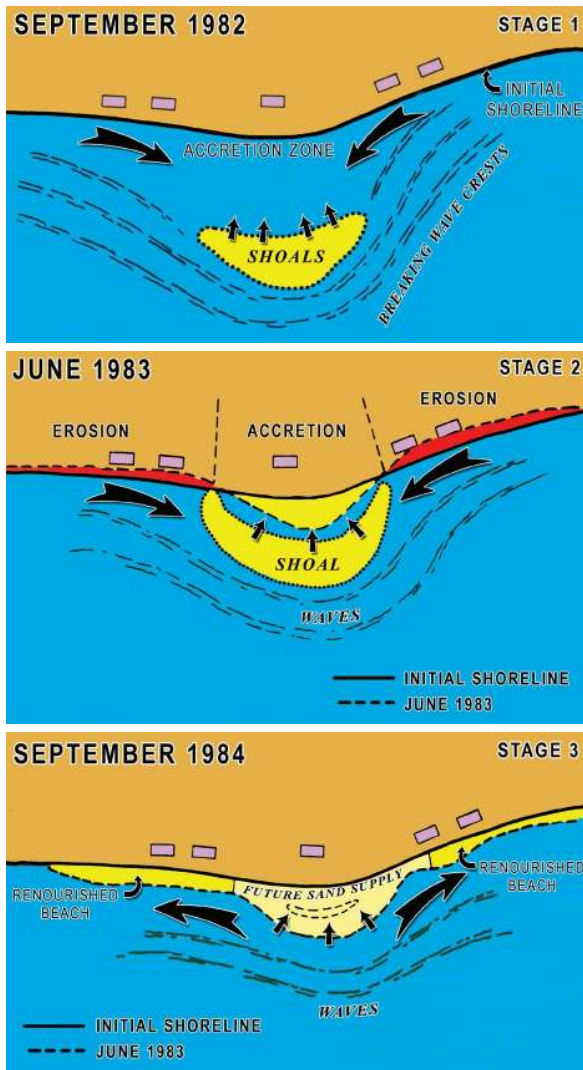


FIGURE 17. The process of shoal bypassing—the episodic transfer of sediment from inlet delta shoals to a beach—accounts for the cycle of erosion and accretion in some settings. Diagram based on actual conditions at Isle of Palms (SC) in the 1980s.

Spit Rotation – Spit formation and growth have been widely linked to longshore transport rates around unstable inlets. Some spits grow aligned with the updrift strand, whereas others tend to rotate landward as they build, changing the shoreline angle (Fig 18). The primary factors influencing spit rotation appear to be (1) the presence of open lagoons or wide channels to accommodate rotation and (2) persistence of washovers on the spit, which locally accelerate erosion with respect to the updrift shoreline. At larger scales, an entire barrier island strandline may rotate as a result of sediment accumulation favoring one end. Century trends for a number of Georgia barrier islands (mixed-energy setting) show seaward growth at the bulbous updrift end and major recession at the downcoast end.

Beaches in Washover Mode – When dunes are absent and backshore elevations low, a portion of littoral sediments become lost from the beach system in storms.



Stage 2 – Shoal bypassing at Isle of Palms, SC, USA

In comparison to adjacent areas with healthy dunes, erosion rates accelerate along beaches in washover mode. Beaches where high dunes (above surge levels) persist, or are artificially maintained, will have comparatively lower net erosion rates because the littoral sediment budget is preserved. The erosion signature will be seen as an accelerated volume change (integrated over the entire profile) within the washover reach. This will be most apparent where a long length of shoreline loses its foredune

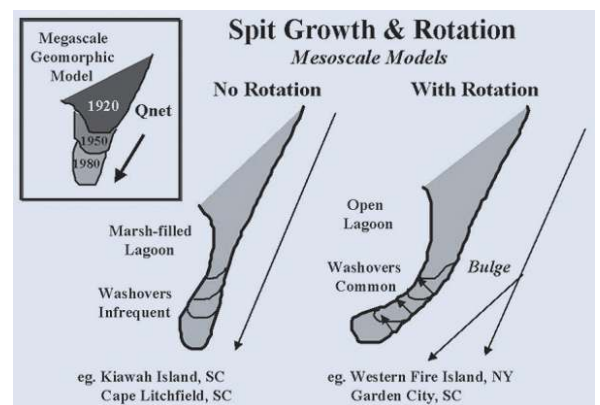
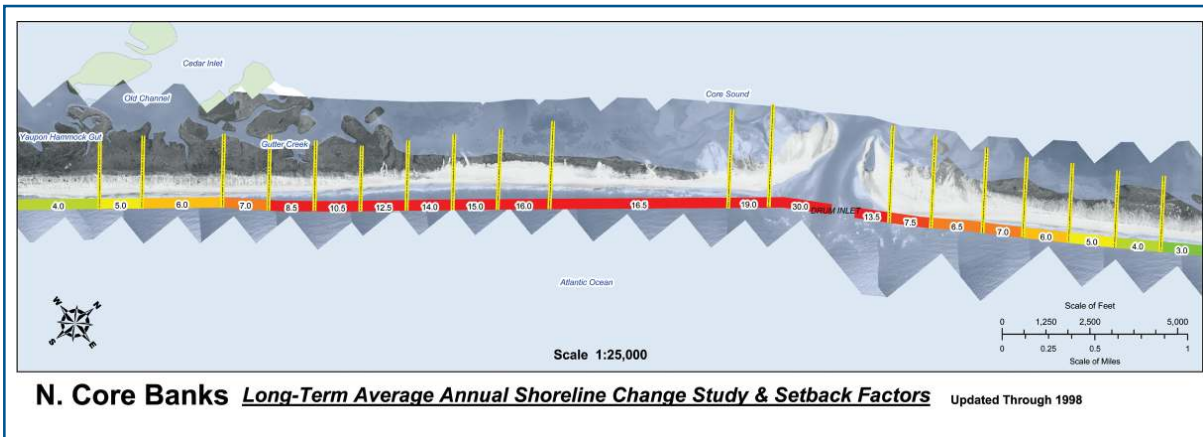


FIGURE 18. Spits form and grow where there is one predominant direction of longshore transport. If the lagoon is mature and filled with marsh, spits will tend to maintain the alignment of the updrift strand. However, if the lagoon is open, there is a natural tendency for the spit to rotate landward as it grows. [Source: Kana 1995]



Washover barrier island – Core Banks (NC) – where 50-year erosion rates range from ~4 ft/yr to 20 ft/yr [Source: NCDENR]

during a major storm and no remedial measures are taken to restore the profile. Examples include parts of Padre Island (TX) and Core Banks (NC).

Depletion of Offshore Shoals – Often updrift of migrating inlets, shoals left from an early tidal or river delta initially hold shoreline salients in place. But over time, the shoal may be depleted as the bottom seeks equilibrium with the surrounding area. Loss of protection from the shoal allows wave energy to focus on the shoreline bulge of the salient. This area becomes a zone of accelerated coastal erosion. There are numerous examples where this process is a factor along the U.S. East Coast, including western Fire Island (NY), south Nags Head (NC), and Debidue Beach (SC) (Fig 19).

There are other signatures at mesoscales that aid in diagnosing erosion problems (Figs 20, 21). Studies have demonstrated fundamental differences between barrier islands in microtidal and mesotidal settings, the former tending to be long and narrow, and the latter tending to be short and wide (Hayes 1979). There are systematic

variations between “reflective” and “dissipative” beaches, the terms referring in this case to the way breaking wave energy is absorbed, and the development of equilibrium profiles along a crenulate beach (Short 1999). Yet, the key point is that most coastal erosion at mesoscales occurs in response to some combination of local processes and sediment supply. Sea-level rise is usually not a significant factor at this scale on a site-specific basis.

Along many coasts, inlets and river deltas control coastal evolution by either withholding or releasing large volumes of sediment to downdrift shorelines. Along drowned river estuaries, lack of a beach cycle leads to one-way sediment movement downslope during storms. Despite the number of ways that coasts erode, many are changing at less than 1 m/yr (Dolan et al 1990, USACE 1995). Some are growing even in the face of rising sea level.

Before presenting some of the defenses against erosion, it is useful to discuss some common misconceptions about coastal erosion.

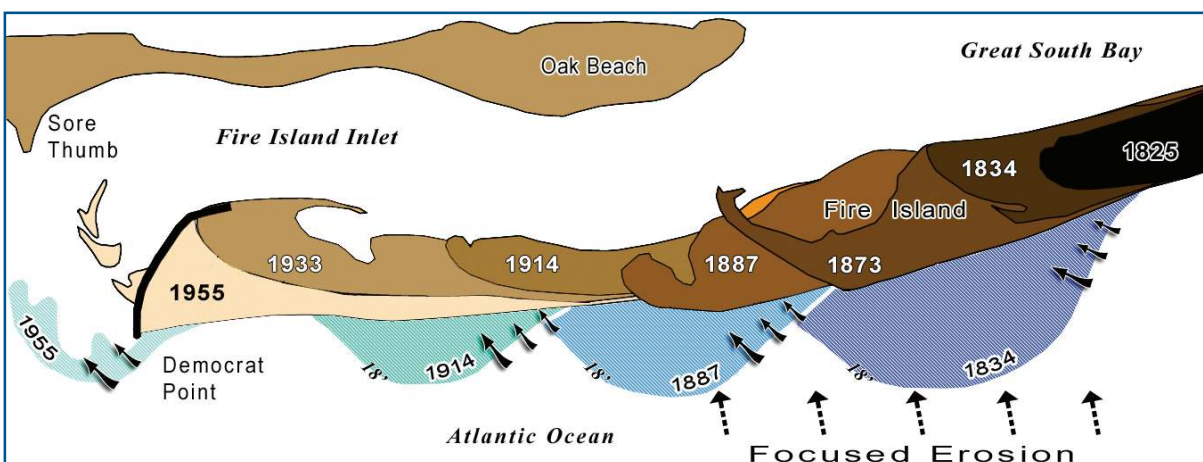


FIGURE 19. If a spit rotates landward as it grows, it leaves a shoreline out of equilibrium with the incoming waves. The broad bulge left updrift becomes a focal point for erosion. In this case from Fire Island (NY), offshore shoals from an earlier inlet have been depleted. Consequently, erosion is now focused along the area of the 1834 and 1887 inlets (the community of Kismet and Robert Moses State Park). [Sources: USACE, CSE]

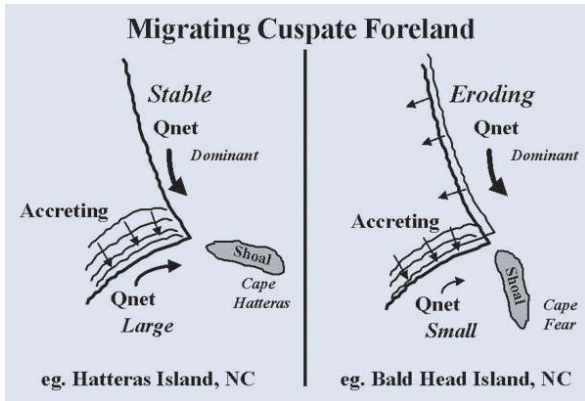


FIGURE 20. Cuspate forelands are coastal promontories where sediment converges under varying wave direction. The shoal orientation at the apex of the foreland influences the rate of erosion and accretion along the adjacent shorelines. [After Kana 1995]

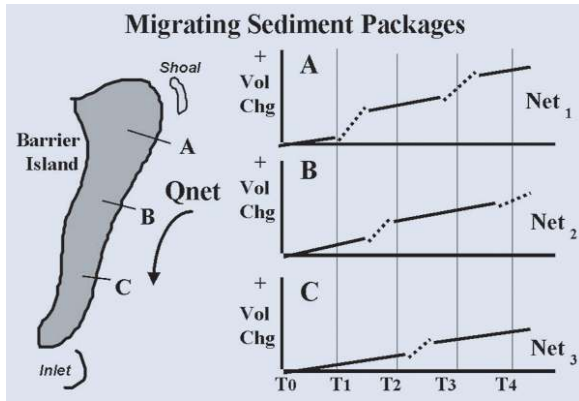


FIGURE 21. After shoals bypass inlets and accrete to the updrift ends of barrier islands, material spreads downcoast in waves. This causes systematic variations in shoreline change rates over time (T). [cf – Bodge 1995]



Garden City (SC) – Groins (center of image) were placed to counter focused erosion along the bulge left from a former inlet.

CONVENTIONAL WISDOM & COMMON MISCONCEPTIONS

By now, it should be apparent there is great variability in the shape of the coast and the processes that transform it. Yet at mesoscales, a surprising amount of shoreline is changing at manageable rates. This is not to say all shorelines should be developed, but rather the vulnerabilities of many areas are not so much due to shoreline retreat.

In the case of America’s West Coast beaches, many of which are anchored between headlands, high backshore elevations limit the runup of surges and waves over the land. Some pocket beaches lose sand to submarine canyons, but many are equilibrium planforms changing at slow rates. Common sense suggests that as long as development is situated well above the highest surge and wave level, and set back some distance from this active zone (and the projected future active zone), the processes of shoreline change will not impact. Other hazards, such as earthquakes and fire, may be of more immediate concern.

Similarly, along some portions of the U.S. East Coast, the backshore is elevated well above the expected 100-year flood elevation [eg – Cape Cod (MA) and Myrtle Beach (SC)], making these areas no more vulnerable to flooding and erosion than inland parcels situated along protected waterways.

Hurricane Damages Also by Wind and Rain

Coastal damages during hurricanes are not always the result of waves and surge. Hurricane Andrew, the fifth most costly on record (Pielke et al 2008), in 1992 destroyed 25,000 homes and businesses in Dade County (FL). Few of these properties were situated on the open coast close to the beach. Instead, most were inland, succumbing to the Category 5 winds (speed >155 miles per hour, or 250 km/hr) associated with the storm.

Much of the property damage in North Carolina after Hurricane Floyd (1999) was due to intense rainfall which dumped over 20 inches (50 cm) on the low-lying mainland dozens of miles from Pamlico Sound, flooding huge areas. The tidal surges originating from the ocean filled estuaries and sounds, thereby impeding the runoff of torrential rains, particularly where “high ground” was only a meter or two above the normal tide limit. One of the ironies of that storm was to meet people who abandoned their flooded mainland homes and sought refuge on Bogue Banks (a barrier island) in second homes perched on land 10 m above the sea.

Yes, living on the open coast is risky, but so is living near the coast.

Hurricane Impacts and Recovery

The impacts of major storms on the coast do not necessarily mean permanent or significant displacement of the shoreline. Hurricane *Hugo* (1989) ranks as one of the top 20 damaging hurricanes of the 20th century (Pielke et al 2008). At Litchfield Beach (SC), the storm caused 25 m (~80 ft) of dune recession, cutting away one of the healthiest foredunes in the state. Emergency scraping, accomplished weeks after the storm, quickly rebuilt a small dune seaward of buildings using sand from the broad, post-storm beach. Sand fencing and grass planting followed so as to jump-start recovery.

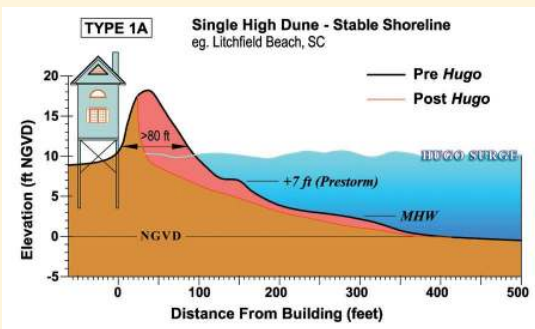


Litchfield Beach (SC) in 1987 before Hurricane *Hugo*. Single foredune reached ~20 ft above sea level with a 100 ft base.

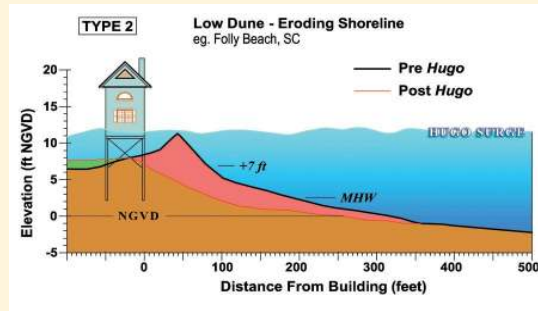


Post *Hugo* washover beach – note negligible damage to building.

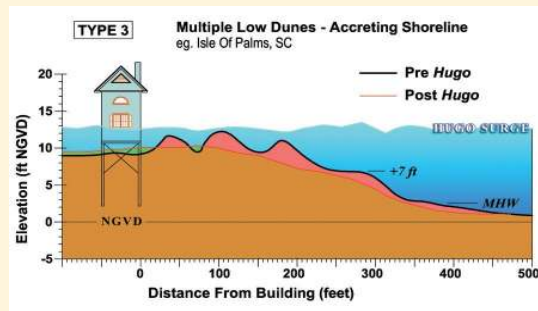
Twenty years later, the emergency dune has grown to its pre-*Hugo* condition (mostly by natural processes). Something similar had occurred along Litchfield Beach after Hurricane *Hazel* in 1954. Recovery in this case stemmed from Litchfield's near-zero shoreline change rate at mesoscales. This erosion and recovery after storms reflects the beach cycle at its finest (cf – Fig 5). There were actually several distinct responses to *Hugo* in South Carolina from beach to beach, depending on the pre-storm condition and long-term erosion trend (Kana 2005).



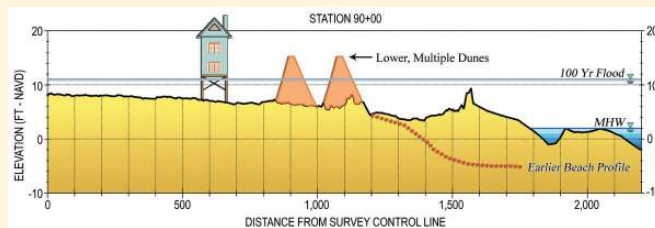
Type 1 Beaches – stable shorelines with a single high foredune (eg – Litchfield Beach) experienced major dune recession but recovered well through emergency scraping and natural profile adjustment.



Type 2 Beaches – eroding shorelines with a low dune or washover profile (eg – Folly Beach and south Pawleys Island) sustained greater damages, loss of sand to washovers, and poor performance of emergency dunes given the sand deficit in the littoral profile.



Type 3 Beaches – accreting shorelines with multiple, low dunes (eg – Isle of Palms) sustained damages despite a large surplus of sand seaward of development. Rapid accretion (along any beach) inhibits formation of high dunes. Backshore areas become stabilized with vegetation before gaining height as more seaward dunes form and intercept wind-blown sand. *Hugo*'s surge overtopped such areas and produced extensive damages to older properties not meeting today's building and elevation standards. Emergency dune construction was successful and long-lasting given the surplus volume in the profile.



FEMA (1986) uses the "540 rule" as a basis for evaluating the level of protection of oceanfront property. Their guideline suggests that there should be at least 540 square feet (equivalent to 20 cy/ft or 50 m³/m) in the dune cross-section above the 100-year flood level to sustain storms such as *Hugo*. Some beaches like Litchfield come close to achieving that condition. But many do not, including places like Sullivan's Island (SC), where 10 ft (3 m) per year accretion has not allowed high dunes to form. In some communities with otherwise healthy building setbacks because of long-term accretion trends, efforts to construct higher dunes conflict with people's desires to maintain views of the ocean or government policies that seek to prevent disturbance of vegetated dune areas even if relief and elevation are insufficient for FEMA-level protection of property.



Bogue Banks (NC) – Thirty five steps to the dry sand beach

Not All Barrier Islands “Roll Over” at Mesoscales

Considering that much of the U.S. East and Gulf Coasts are chains of barrier islands, coastal erosion is often associated with that landform. The conventional wisdom suggests these “ephemeral islands” are doomed to overtopping in storms and destruction under rising seas. The image conveyed is that of a bulldozer tread where storms overtop the barrier, drive sand inland, and roll the island into the lagoon. Back-barrier marshes are temporarily buried by overwash only to emerge as outcrops on the ocean-side beach some time later. This image is compelling because it can be seen happening today in places like Louisiana, Bolivar Island (TX), or Edingsville Beach (SC) where there was a group of “planters cottages” in the 1880s until a hurricane destroyed (or moved) every building (cf – Fig 6). Sallenger (2009) brings to life what happens when catastrophic storms impact low barrier islands.

Some of the highest coastal erosion rates (>3 m/yr) occur on “washover” barrier islands like Core Banks (NC) or

Assateague Island (MD) (on which much of the conventional wisdom regarding “barrier island rollover” is based). Yet, the majority of developed barrier islands do not fit this simple model when considered at century time scales. Bogue Banks (NC), where some people sought refuge after Hurricane Floyd, is over a kilometer wide with dunes reaching 15 m heights. There have been no breaches of the 40-km-long island in the last hundred years, and erosion rates average well under 1 m/yr (CSE 2007). Likewise, Long Beach Island (NY), Atlantic City (NJ), and Miami Beach (FL) are urban barrier islands with no known history of cross-barrier overwash or breaching during the 20th century.

Galveston (TX) was overwashed and its community largely destroyed in the great hurricane of 1900 (Larson 1999). After the storm, the land was pumped up via dredge spoil, raising the core of the island well above most (but not all) hurricane surge levels. Buildings were raised and a seawall was constructed to further inhibit washovers. [Note: The same cannot be said for the east end beaches (such as Jamaica Beach) on Galveston Island, which were not built up after the storm, or for neighboring Bolivar Island (a low barrier with little relief) which was devastated by Hurricane Ike in 2008.]

Another barrier island that has not washed over since the 1920s is Jones Beach (NY). One of the great recreational beaches in the world, Jones Beach owes its persistence to Robert Moses, New York’s builder of parks and parkways, who arranged to pump 30 million cubic meters from Great South Bay along the central spine of a chain of washover barriers. The parkway on top of this spine sits above flood levels and provides access for millions of beachgoers every summer.



FIGURE 22. The eastern end of Kiawah Island (SC) (2008) where shoal-bypassing events between 1990 and 2005 added 5 million cubic yards to the island and created a 3-mile-long barrier beach and lagoon system. Episodic events such as this illustrate how rapidly barrier islands can form in the presence of an ample sediment supply. [Source: CSE 2009]



Indian Beach – Bogue Banks (NC) – after Hurricane Floyd (1999) – top of escarpment is >20 ft above MSL

Barrier Islands Higher than Mainland Shores

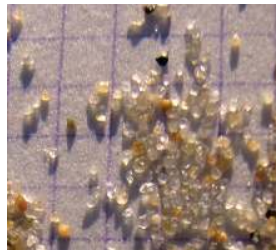
This is not to say that most barrier islands will never wash over. Certainly, if sea levels rise faster than the models predict, this will become commonplace. Yet, barrier islands are, by their very nature, landforms which build vertically under the action of waves and winds. If sea level rises and there is sufficient sandy sediment in the littoral zone and back barrier area (above the DOC), the island will reconstitute itself. Given sufficient sediment supply, barrier islands can form rapidly, even in the face of rising seas (see Fig 22). Of, perhaps, more immediate concern for coastal zone management are those low-lying interior lands that are not exposed to the building processes of ocean waves. Potentially much more development will be at risk sooner along sheltered estuaries where the land is only a meter or so above present high tide. Mastic Beach (NY) is just such a community vulnerable to sea-level rise before its protective barrier, Fire Island, is likely to be washed over in many places.

A misconception regarding barrier-island retreat is that the average slope of the inner continental shelf will control the retreat distance. Some scientists assume that if the average slope of the shelf is .001 (1 m rise over 1,000-m distance, which is a characteristic value), a 1-m rise in sea level will translate to a 1,000-m shoreline recession. While this may be the case over long geologic time scales, empirical evidence shows it is not the case at century time scales in most areas. As Bruun (1962) demonstrated, sea-level rise causes the beach and barrier island to be displaced landward and upward in proportion to the slopes in the active surf zone (which are much steeper than the inner shelf). Therefore, if the slope of the beach is ~1 on 30 (a typical value), a 1-m sea-level rise would produce ~30-m shoreline retreat. Remember, beaches tend to equilibrate when their profiles achieve the “normal” healthy volume for the setting (cf – Figs 7 and 15). Yes, many other factors modify this result, but not to the degree that the shoreline soon migrates 1 kilometer (km) inland. This becomes important in the debate about solutions.

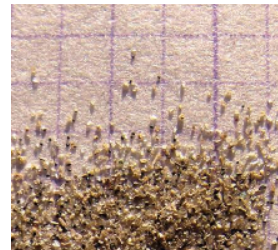
Muddy vs Sandy Barrier Island Coasts

One final misconception about the response of the coast to erosion and rising seas is related to sediment quality. Here again, Louisiana provides an object lesson. Relative sea-level rise near the mouth of the Mississippi River was ten-fold greater than the global average rise during the 20th century. Barrier islands are disintegrating, marshes are drowning, and the land is retreating as fast as anywhere in the world. “As Louisiana goes, so will the US East Coast barriers if sea-level rise accelerates” as some worst-case scenarios predict. But this ignores a key difference between Louisiana barrier islands and most East Coast barriers. Mud and very fine-grained sand are the dominant sediments along the Louisiana coast, whereas fine- to coarse-grained sand dominates along the East Coast.

A typical mean grain size of a Louisiana beach is 0.1 mm, whereas a North Carolina beach is 0.3 mm. This may not seem like a big difference, but it translates to much gentler slopes in Louisiana compared with North Carolina.



Large grain sand (0.4 mm)



Fine grain sand (0.1 mm)

If one considers the cross-section of a barrier island to be similar to a prism, the Louisiana barrier will be very broad and flat, whereas the North Carolina barrier will be much narrower at the base and steeper along the sides. Louisiana’s outer beaches may equilibrate at slopes of 1 on 100, whereas North Carolina’s are closer to 1 on 20. Few barrier islands in Louisiana grow more than 1–2 m above high water, while many North Carolina barrier islands have dunes over 5–10 m high (Fig 23). Simple geometry of each cross-section shows that Louisiana needs much more muddy sediment to build a broad base before a barrier of 0.1 mm sand can accumulate on top of the mass above the high-tide level. Louisiana’s critical need is for more coarse sediment at the coast, which it does not have. North Carolina beaches tend to be founded on coarser sands (0.2 mm in the underwater zone) and composed of 0.3–0.5 mm grain sizes in the dry-beach zone. Not surprisingly, then, Louisiana’s barriers are not keeping pace with local sea-level rise, despite valiant attempts to maintain them (Fink and Khalil 2005). South Carolina, where littoral sediments average 0.2 mm, contains examples of barrier beaches that have emerged from the ocean and have established themselves in a decade (Kana 2002, CSE 2009).

The point of this is to say that the Louisiana response to rapid sea-level rise is not necessarily the best model for

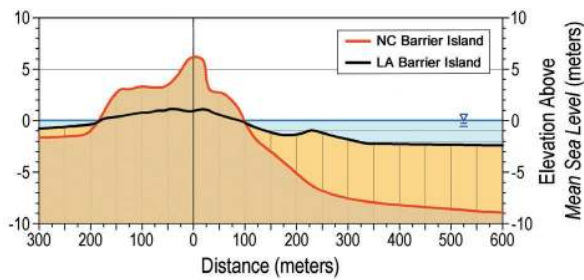


FIGURE 23 Characteristic cross-sections of barrier islands in North Carolina and Louisiana. Coarser sandy sediments lead to a steeper profile and less volume in the base of North Carolina barrier islands compared with the muddy barrier islands of Louisiana which are founded on fine-grained sediments of the delta. Note low relief of the Louisiana cross-section. Ocean is to the right.

many developed barrier islands where sediments are coarser. Just as road builders prefer a base of coarse sand which drains well, barrier islands will form readily in the presence of sandy sediments with higher porosity (cf – Fig 22). They will persist longer under the action of waves, winds, and tides if dunes grow above the storm-surge levels. Washover frequency will lessen along with recession rates because the littoral budget will be conserved seaward of the foredune. Storms may erode the dune, leaving escarpments, but the beach cycle will rebuild the profile using eroded sediment. As long as a healthy profile volume and sufficient wave energy exist in the littoral zone, a barrier beach will be maintained. In contrast, low barrier islands of Louisiana will continue to lose a portion of their littoral volume to the lagoon whenever the islands wash over or breach. This extra loss will yield higher recession rates. How one manages shoreline erosion and development along a more frequently overtopped barrier island, like Bolivar Island (TX), should probably differ from efforts along slowly eroding, high islands, such as Bogue Banks (NC).

COASTAL EROSION DEFENSES

Defense against coastal erosion generally takes two forms—hard solutions involving shore-protection structures and soft solutions involving manipulation of sediment supplies or political controls restricting coastal development. An intermediate defense can also be classified whereby hard, sediment-retaining structures (such as groins, breakwaters, and jetties) are combined with nourishment.

Hard Solutions

When fixed, permanent development along sedimentary coasts is necessary, such as ports and harbors, shorelines are armored by means of bulkheads, revetments, or seawalls. These shore-parallel structures are designed to stabilize the backshore, prevent encroachment of the sea, and retain sediments behind the structures. Some are low cost and short-lived (eg – geotextile bags that are pumped full of sand and stacked at the edge of the shoreline, Fig 24) while others have design lives exceeding 100 years (eg – massive stone revetments). The principal design factors

are the size of waves and height of tides at a site. Because wave power increases by the height-squared of the wave, shore-protection structures along high energy coasts must be massive to withstand the forces (ASCE 1994).



FIGURE 24. Failed, low-cost shore protection. Once punctured, large sand bags leak their contents and are no longer effective.



Massive quarry-stone seawall protecting private property (1985) before restoration of the beach – Seabrook Island (SC) [see cover photos]

The problem with hard solutions is that they do not preserve the littoral profile. As noted earlier, coastal erosion (ie – loss of sediment volume in the littoral zone) continues in the presence of structures (unless the shoreline segment is accreting over the long term). As sediment is lost, the beach profile is lowered at the structure, thus allowing higher waves to impact. This complicates the design because it means wave conditions will change over time as the profile continues to erode. Large civil-works projects, such as port facilities, generally take this into account. However, small private projects, such as revetments for waterfront homes, often ignore this factor, opting for the lowest cost structure needed for the wave conditions at the time of installation.

In the 1960s and 1970s, many oceanfront property owners in the U.S. constructed private seawalls to protect property, only to find a few years later that they were inadequate. In many cases, the original structure required frequent repairs and upgrading to maintain its structural integrity.

The cost of shore-parallel structures spans a wide range because of variations in scale and exposure to waves. Along exposed ocean coasts, the typical range in U.S. dollars in

the year 2010 was \$1,000 to \$5,000 per meter of shoreline. A useful way of evaluating costs is to compare the average annual cost of a shore-protection structure with the value of the property being protected. If the annual cost is less than 1-2 percent of the property value (ie – comparable to annual property taxes in the U.S.), the investment may be appropriately proportional over the typical life of buildings. But when coastal armoring (including annual maintenance costs) exceeds ~5 percent of property values, the economics start to become unfavorable. Another factor that should be considered is the cost of losing the beach if erosion continues.

Increasingly, it is recognized that backshore armoring may save upland development but at the expense of the littoral profile. Along beach communities, this often leads to loss of recreational opportunities and lower property values compared with communities that maintain healthy beaches. Some states, including North Carolina and South Carolina, have banned new seawalls in an effort to preserve recreational beaches.

Semi-Soft Solutions

Semi-soft solutions utilize hard structures to trap and retain sediment while leaving the adjacent shoreline natural. Nourishment is sometimes mandated as a condition of building sand-retaining structures such as jetties, groins, and breakwaters. As described in the section on signatures of coastal erosion, these artificial structures are often a cause of downcoast erosion. At the same time, they offer proven erosion control along upcoast segments, as the Dutch have demonstrated for hundreds of years. Groins are no longer favored, but in some settings, they represent the most feasible method of slowing erosion rates and holding the shoreline in place (see photo, pg 2). This is particularly true around the margins of tidal inlets or where a relatively short segment of beach such as a spit is losing sand more rapidly than upcoast areas (ASCE 1994). Today, best management practice discourages construction of groins and breakwaters unless it can be shown that they will not adversely affect downcoast areas, or that mitigation such as nourishment is included. Sites where this applies include the downcoast ends of littoral cells, reaches isolated from adjacent areas by major channels, headlands, or other littoral barriers, and areas where erosion rates are exceedingly high (cf – Fig 25) A rule of thumb is that such structures are generally not economic where erosion rates are less than 1 m/yr because of the high capital cost as well as aesthetic concern. Soft solutions, such as nourishment and wider development setbacks, are favored in those cases.

Soft Solutions

Beach Nourishment (ie – the addition by artificial means of non-littoral sediment along the shoreline) is the primary soft solution for coastal erosion. It is the only solution that at once preserves the recreational beach and protects upland development (NRC 1995). Other soft solutions

related to nourishment involve manipulation of sand supplies in the littoral zone, as follows.



FIGURE 25. Severe, localized erosion adjacent to a tidal inlet (Ocean Isle Beach, NC). Net longshore transport is into the inlet. A terminal groin positioned downdrift of these houses (ie – middle of photo) is probably the most cost-effective way to hold the beach in place with a secondary benefit of reducing shoaling in the channel. Once filled to capacity, properly designed groins will bypass excess sediment.

Beach Scraping – This is a practice whereby sediment from one portion of the profile (usually near the mid-tide line) is moved mechanically to another (usually the base of an eroded dune). Not technically nourishment because it does not add to the sediment supply, scraping is often performed after storms in an attempt to accelerate the natural beach cycle and provide temporary protection to threatened property. Typical U.S. cost for emergency scraping in the year 2010 was around \$15–\$30 per meter (~\$5–\$10/ft) of shoreline. This assumes movement of ~5–10 m³/m (2–4 cy/ft).



Beach scraping via bulldozer

Transfer from Accretion Zones – Similar to beach scraping, sediment is mechanically transferred in the longshore direction from zones of accretion (or healthier sections) to eroding sections. For example, a sand-spit at the downdrift end of a beach may be accumulating



Pan earth movers excavating low-tide terrace in accretion zone for transfer to erosion zone

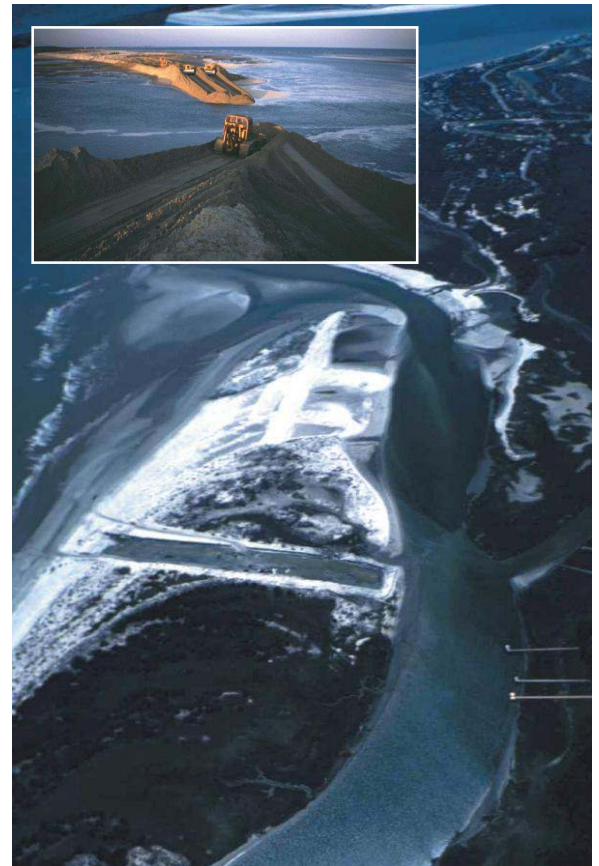
sediment derived from the updrift section. Excess sediment may be “borrowed” from the spit and recycled back to the eroding areas. This practice is advisable only where the history of spit growth is well known and careful monitoring is performed to track the regional sediment budget. Typical year 2010 U.S. costs for transfer from accreting areas were \$1.50 per cubic meter per kilometer (m^3/km) transfer distance. In some cases involving large volumes and



Excavating a tidal inlet through an incipient barrier spit

moderate distances, the transfer can be performed more economically by hydraulic dredge. So to add $25 m^3/m$ (10 cy/ft) using an accreting area 3 km (~2 miles) away, the restoration cost would be (~)\$100–\$125/m (~\$30–\$40/ft).

Inlet Relocation/Channel Realignment – This method is applicable along shorelines bounded by unstable inlets. It seeks to move channels, which may be encroaching directly on the littoral profile, away from the area. A secondary benefit is the release of sediment from the abandoned inlet. Inlet relocation allows the shoals in tidal deltas to move onshore and naturally nourish the downcoast shoreline. There is little experience with this innovative method but in the case of Captain Sams Inlet (SC), two relocation events (1983 and 1996) each provided over 1,000,000 m^3 to an eroding beach at a cost of less than US\$0.5/ m^3 (Kana 1989). The resulting nourishment widened a 3-km section



Captain Sams Inlet Construction – Seabrook Island, SC, USA

of beach by over 300 m. [See cover photos.] These projects were exceedingly cost effective and environmentally friendly (NRC 1994) because they were executed quickly by land-based equipment (low mobilization costs) and the volume moved by trucks (~150,000 m^3) was dwarfed by the volume eventually moved by natural processes.

Composite and Jurisdictional Solutions – Much management of coastal erosion entails jurisdictional solutions. As some researchers have stated, erosion is only a problem where there is human development. Therefore, if building is prohibited in erosion-prone areas, the problem will be inconsequential. Unfortunately, eliminating development along eroding coasts is not simple. People have been drawn inexorably to the shore for centuries and are often unwilling to abandon retreats that have been passed down for generations, even if erosion remains a constant threat.

Jurisdictional solutions usually begin with delineation of flood and erosion hazard areas. Lines are drawn on maps showing where the potential problem areas lie. This idea is simple in concept but difficult to put into practice because it bumps into ownership and property-rights issues. Historical practice was to place development some distance back from the coast, using vegetation and other indicators

the most famous beach in the world in 1900. By 1923, its beach had been lost to erosion, replaced by seawalls, jetties, and groins. In that year, over 2,000,000 m³ of sand were pumped onto Coney Island's 8 km of shore, restoring the beach to its former glory. Other than some minor nourishment in the 1940s and 1960s, Coney Island held most of its sand for decades, partly



thanks to jetties and groins. In 1996, another project the size of the first one was completed. Great beach resorts only stay that way if there is a beach. From Coney Island to Miami Beach, Sunset Beach to Waikiki, beach nourishment has preserved some famous beaches around the world.

Despite many successes, there is still controversy about nourishment with some arguing that it is impermanent and a waste of money, while others claiming it will solve all coastal erosion problems. To view the solution objectively, one must weigh the economics of nourishment against the other alternatives, including property abandonment and relocation.

The cost of nourishment is fundamentally linked to two factors: (1) the rate of erosion and (2) the cost of importing new sand. Both of these factors can vary by an order of magnitude from site to site. A beach eroding at 1 m/yr will require much less nourishment sand than another eroding at 10 m/yr. Similarly, sand costing \$1/m³ is a better value than \$10/m³ sand. One site may cost \$5/m/yr to maintain a particular littoral volume while another site may cost \$100/m/yr. Using an assumed average of \$50/m/yr for a 10-km segment of beach, it would cost about \$5 million every ten years just to keep pace with erosion. Yes, in the aggregate, this is expensive, but how does it compare with alternatives?

Doing nothing entails the costs of land loss, property abandonment, and diminishment of the local tax base. Assume in the example above that the average erosion rate is 1 m/yr. After 50 years, 50 m of oceanfront will be lost along with buildings and infrastructure in that

zone. Now, again using some typical 2010 values, assume the oceanfront lots are 50 m deep and extend 50 m along the oceanfront. There will be ~200 properties along the 10-km beach. In 1960, one could purchase such property in the Carolinas or Florida for as little as \$10,000. Today (2010), there are few oceanfront properties of this size that sell for less than \$200,000. In many places from Hilton Head Island (SC) to Rehoboth Beach (DE), such properties command over \$1 million. Property tax rates, even if only 1-2 percent of assessed property values, generate significant revenues to local communities. At \$500,000 per parcel, the potential 50-year loss (before adjusting for inflation) would be over \$100 million, or about four times the cost of maintaining the beach (in this example).

Relocation of buildings is only practicable if lots are very deep or vacant land is readily available in the beach community. Rarely is this the case. Costs of relocation are generally comparable to costs of abandonment (USACE 2010).

Beach nourishment has become the default solution at mesoscales largely for one reason—oceanfront property values have risen many times faster than the cost of nourishment over the past century. As the graph shows, sand delivery cost has risen about tenfold since 1950. By comparison, oceanfront property values have risen about 1,000-fold in the same time frame. As long as these two lines keep diverging, beach nourishment will remain more cost effective than property abandonment. And if the investment includes beach widening and dune enhancement (following the example of Myrtle Beach, SC), the net result will be improved building setbacks and reduced storm damages.

CLOSING THOUGHTS

As logical as scientists and engineers try to make the problem of coastal erosion and formulate cost-effective solutions, the debate often breaks down into special interest camps with champions pitted against opponents. This booklet is not expected to change that. But drawing on experience and considering the coastal economy at risk, more objectivity is needed. And that begins with better and more accurate measurements of the sand volume lost each year along all developed shorelines. Communities that do not measure their beach losses can hardly know what it will cost to restore and maintain the beach.

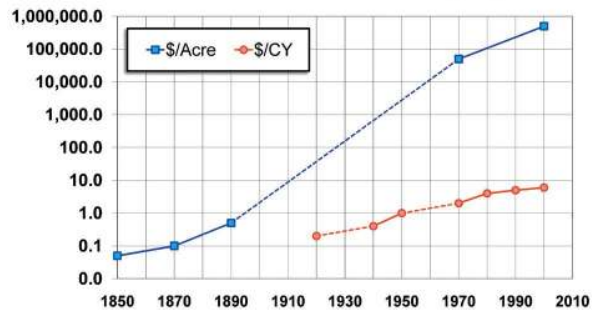
There is far greater experience today with soft solutions than there was in the 1970s. Many more projects are lasting a decade or so before it is time to renourish or to relocate a migrating inlet. Poor site planning, overdevelopment, and laissez-faire structural solutions by individual property owners exacerbated the erosion problem during the 1970s. The availability of flood insurance perhaps, compounded the problem and fostered a trend of

larger replacement houses after storms did their damage. Government officials charged with regulating coastal development are still struggling to balance the rights of property owners with the needs of public safety, recreation, and environmental quality.

Accelerated sea-level rise, if it indeed occurs this century, will obviously affect the economics of beach maintenance and coastal damages in storms. It is difficult to convince a life-long resident of Cape May (NJ) to abandon a family home if we do not abandon places like the 9th Ward in New Orleans (LA), which sits 3 m below sea level. Much of the coast is now held in trust as national seashores, wildlife preserves, and related conservation zones. This is a great legacy of the 20th century. Meanwhile, economics should drive most decisions regarding remaining areas that are developed. Further complicating the issue is controversial subsidies for reconstruction or beach building.

Coastal property owners can only deal with sea-level rise and chronic erosion by banding together as a community and implementing regional rather than individual solutions. One person can build a fortress around their property, but it takes a large group of people to build and maintain a beach. Nourishment, as well as any semi-soft solution, lasts longer and is more cost effective performed on a community-wide scale. Nourishment longevity increases roughly with the square of the project length (Dean 2002), so a 10-km project will tend to last four times longer than a 5-km project. Some of CSE's longest-lasting projects were 13–20 km long.

**Bogue Banks, NC Property Values - (\$/Acre)
And U.S. Nourishment Costs - (\$/cy)**



Most U.S. East Coast beaches sustained a (roughly) 1-ft (30-cm) rise in sea level in the 20th century. Some accreted, many eroded, but development generally expanded. If sea-level rise doubles, the accommodation of that rise will have to be made in half the time or about 50 years. And if sea-level rise increases fivefold in the same time (as some believe), the adjustment will have to occur in barely 20 years. Before sea level rises 5 ft (1.5 m), it has to rise 1 ft, then 2 ft, and so on. As simple as this statement is, society will be wise to monitor tides at the coast. This is the clear and convincing evidence needed to plan for the future. A 1-ft rise in 20 years will be obvious in the event and, in areas like South Carolina's Lowcountry, will likely produce more immediate effects along interior areas. Flooding around estuaries and low causeways may actually drive development decisions before erosion along our beaches.



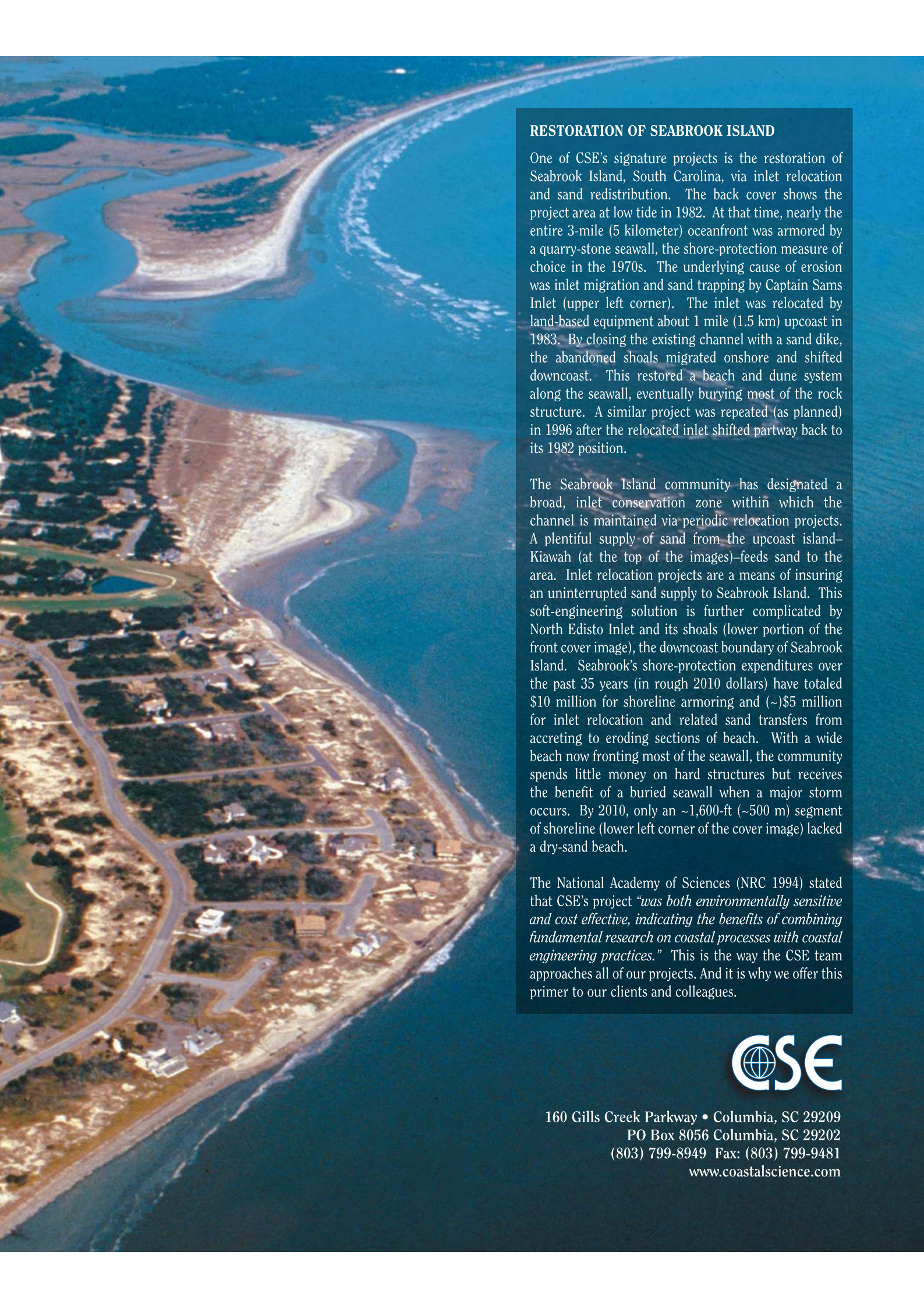
CSE has planned and directed more than 35 beach restoration projects since 1984, covering 70 miles of shoreline. The present value of oceanfront properties enhanced by these and related federal beach nourishment projects is well in excess of US\$10 billion.

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RESTORATION OF SEABROOK ISLAND

One of CSE's signature projects is the restoration of Seabrook Island, South Carolina, via inlet relocation and sand redistribution. The back cover shows the project area at low tide in 1982. At that time, nearly the entire 3-mile (5 kilometer) oceanfront was armored by a quarry-stone seawall, the shore-protection measure of choice in the 1970s. The underlying cause of erosion was inlet migration and sand trapping by Captain Sams Inlet (upper left corner). The inlet was relocated by land-based equipment about 1 mile (1.5 km) upcoast in 1983. By closing the existing channel with a sand dike, the abandoned shoals migrated onshore and shifted downcoast. This restored a beach and dune system along the seawall, eventually burying most of the rock structure. A similar project was repeated (as planned) in 1996 after the relocated inlet shifted partway back to its 1982 position.

The Seabrook Island community has designated a broad, inlet conservation zone within which the channel is maintained via periodic relocation projects. A plentiful supply of sand from the upcoast island—Kiawah (at the top of the images)—feeds sand to the area. Inlet relocation projects are a means of insuring an uninterrupted sand supply to Seabrook Island. This soft-engineering solution is further complicated by North Edisto Inlet and its shoals (lower portion of the front cover image), the downcoast boundary of Seabrook Island. Seabrook's shore-protection expenditures over the past 35 years (in rough 2010 dollars) have totaled \$10 million for shoreline armoring and (~)\$5 million for inlet relocation and related sand transfers from accreting to eroding sections of beach. With a wide beach now fronting most of the seawall, the community spends little money on hard structures but receives the benefit of a buried seawall when a major storm occurs. By 2010, only an ~1,600-ft (~500 m) segment of shoreline (lower left corner of the cover image) lacked a dry-sand beach.

The National Academy of Sciences (NRC 1994) stated that CSE's project "*was both environmentally sensitive and cost effective, indicating the benefits of combining fundamental research on coastal processes with coastal engineering practices.*" This is the way the CSE team approaches all of our projects. And it is why we offer this primer to our clients and colleagues.



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