Coastal Impacts Due to Sea-Level Rise

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Key Words

barrier islands, tidal inlets, salt marsh, wetlands, inundation, estuaries, equilibrium slope

Abstract

The Intergovernmental Panel on Climate Change (2007) recently estimated that global sea level will rise from 0.18 to 0.59 m by the end of this century. Rising sea level not only inundates low-lying coastal regions but also contributes to the redistribution of sediment along sandy coasts. Over the long term, sea-level rise (SLR) causes barrier islands to migrate landward while conserving mass through offshore and onshore sediment transport. Under these conditions, coastal systems adjust to SLR dynamically while maintaining a characteristic geometry that is unique to a particular coast. Coastal marshes are susceptible to accelerated SLR because their vertical accretion rates are limited and they may drown. As marshes convert to open water, tidal exchange through inlets increases, which leads to sand sequestration in tidal deltas and erosion of adjacent barrier shorelines.

INTRODUCTION

Many scientists consider global warming–forced climatic change as the most serious environmental threat facing the world today (IPCC 2007). Global warming has the potential to affect many humans dramatically and adversely as a consequence of both natural and anthropogenic changes to temperature, precipitation, sea level, storms, air quality, ecosystems, and other climatic conditions.

Sea-level rise (SLR) poses a particularly ominous threat because 10% of the world's population (634 million people) lives in low-lying coastal regions within 10 m elevation of sea level (McGranahan et al. 2007). Much of this population resides in portions of 17 of the world's 30 largest cities, including Mumbai, India; Shanghai, China; Jakarta, Indonesia; Bangkok, Thailand; London; and New York (**Figure 1**). The population of many of the Asian cities will likely continue to increase as ports

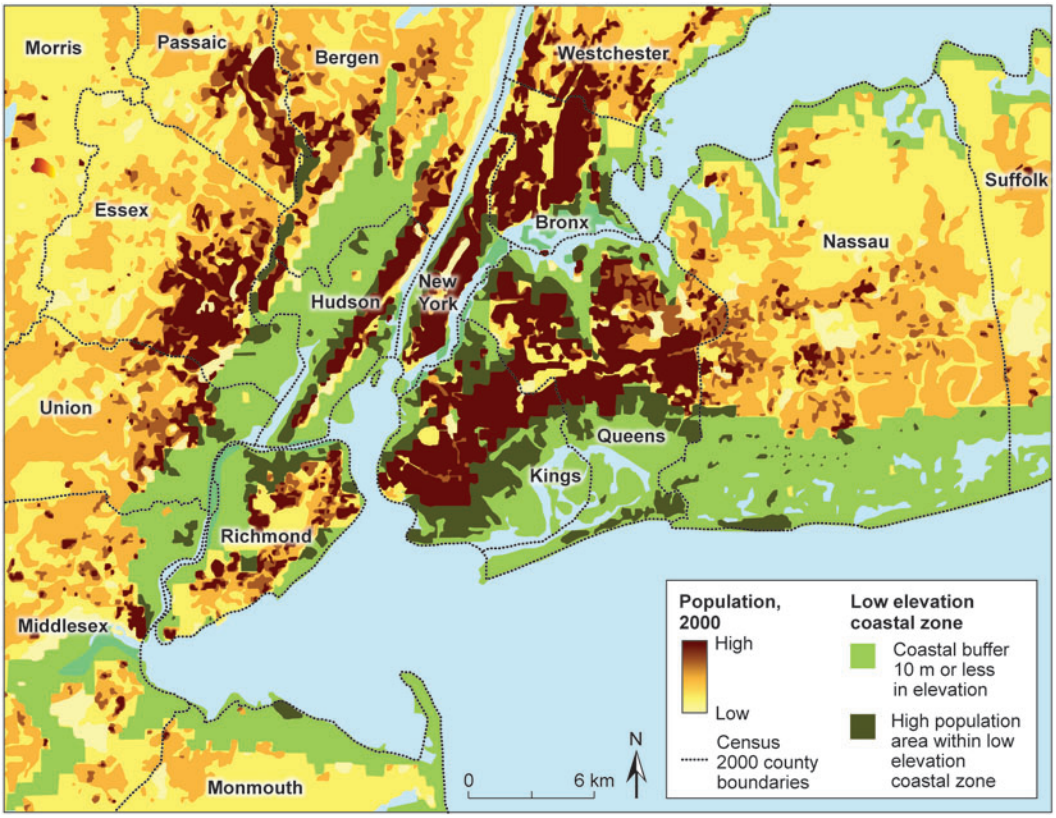


Figure 1

Map of Metropolitan New York showing areas of high population and vulnerability to coastal flooding due to their low elevation (from **http://www.earthinstitute.columbia.edu/news/2006/story05-12-06b.php**). The funneling effect of the New Jersey and Long Island shores produces enhanced storm surge heightening during the passage of hurricanes that make landfall west of New York (Coch 1995).

and work forces expand to keep pace with economic globalization and increasing shipping traffic (McGranahan et al. 2007). Between 1980 and 2003, the population of 672 coastal counties in United States increased from 120 million to 153 million people (to 53% of the total population) and this number is expected to rise to 160 million people by 2008 (Crossett et al. 2004). In 2003, coastal counties in the United States accounted for 23 of the 25 most densely populated counties.

In addition to inundating low-lying coastal areas, rising sea level increases the vulnerability of coastal regions to flooding caused by storm surges, tsunamis, and extreme astronomic tides. As sea level rises, storms of a given magnitude reach higher elevations and produce more extensive areas of inundation.

Likewise, storm surges of a given height have greater recurrence intervals. Rising sea level causes natural thresholds to be exceeded more frequenty, which, in turn, leads to greater occurrences of waves breaking over seawalls, flood waters overtopping levees, and storm surges overwashing and breaching barriers. In areas affected by tropical storms, warmer ocean surface temperatures may exacerbate these conditions by increasing the magnitude of storms (Webster et al. 2005, Emanuel 2005). The recent loss of life and destruction of property in the northern Gulf of Mexico due to hurricanes Katrina and Rita in 2005 underscore the vulnerability of coastal regions to storm surges and flooding. The potential loss of life in low-lying areas is even more graphically illustrated by the 1970 Bhola cyclone that traveled northward through the Bay of Bengal producing a 12-m-high wall of water that drowned a half million people in East Pakistan (now Bangladesh) (Garrison 2005).

The long-term association of population centers with lowland coastal regions dates back to early civilizations when people congregated at river mouths and estuaries because of abundant and accessible food sources (Stanley & Warne 1997, Kennett & Kennett 2006). A recent theory now ties the emergence and rapid expansion of many of the initial complex societies to a slowing rate of SLR (Day et al. 2007, Horton 2006). These authors suggest that the rapid growth of these complex societies occurred when rising sea level began decelerating approximately 7000 years BP following deglaciation. They argue that prior to that time, sea level rose too quickly [1 m century^{-1} (Fairbanks 1989)] to permit coastal communities to become permanently established and to prosper. Prior to 7000 years BP, shorelines along low-gradient coastal zones, such as delta plains, retreated at a rate of approximately 1 km century^{-1} (Day et al. 2007). The commonality in response of early civilizations and recent inhabitants of coastal regions, such as the rapidly subsiding Louisiana lower delta plain, the disappearing islands in Chesapeake Bay, and other abandoned lands to the encroaching sea, emphasizes the degree to which SLR has and continues to influence human populations.

Many of the present ills associated with rising sea level represent the cumulative effects of processes that have been ongoing for many decades and perhaps centuries. In fact, these effects may be related to other natural and anthropogenic factors in addition to SLR, such as reduced or exhausted sediment supplies and human actions. Despite the possible influence of these other factors, SLR may still have served as the major forcing agent in promoting long-term erosion of coasts worldwide (Leatherman et al. 2000, Pilkey & Cooper 2004). As acknowledged in the recent IPCC (2007) report, a growing number of tide gauge and field studies demonstrate that the rate of SLR began increasing between the mid-nineteenth and mid-twentieth centuries, (Donnelly 2006, Donnelly & Bertness 2001, Gehrels 1999, Nydick et al. 1995) and recent tide gauge data suggest that since 1993, the rate of SLR has increased to 3 mm year^{-1} (Church & White 2006). Thus, many of the impacts of accelerating SLR can be generalized as worsening widespread existing conditions. For example, flooding lowlands, beach erosion, saltwater intrusion, and wetland loss are all processes that have been ongoing along coasts for centuries and have been widely recognized for many years (Bird 1993, Leatherman 2001).

In addition to increased flooding and greater storm impacts to coastal communities in many low-lying regions, accelerated SLR will dramatically alter sandy beaches and barrier island coasts. These impacts go beyond simple inundation caused by rising ocean waters, and involve the permanent or long-term loss of sand from beaches. The loss results from complex, feedback-dependent processes that operate within the littoral zone, including onshore coastal elements (e.g., the nearshore, beachface, dunes, tidal inlets, tidal flats, marshes, and lagoons). Sediment budget analyses have shown that nearshore, tidal deltas, capes, and the inner continental shelf can serve as sediment reservoirs (Komar 1998). Long-term beach erosion may increase due to accelerated SLR and may eventually lead to the deterioration of barrier chains such as those along the U.S. East and Gulf coasts (Williams et al. 1992, FitzGerald et al. 2007), Friesian Islands in the North Sea, and the Algarve coast in southern Portugal. Barriers protect highly productive and ecologically sensitive backbarrier wetlands as well as the adjacent mainland coast from direct storm impacts and erosion. Moreover, barriers support residential communities and a thriving tourist industry. An estimated $3 trillion are invested in real estate and infrastructure on the barriers and mainland beaches along the East Coast of the United States (Evans 2004). A single 7-km-long barrier in North Carolina, Figure Eight Island, has a tax base of more than $2 billion (W. Cleary, personal communication). Tourism plays a major part in the economies of developing countries, and the success of this industry depends on the vitality of the beaches.

Determining the socioeconomic impacts of SLR on coastal areas consitutes one of this century's greatest challenges (Titus & Barth 1984, Titus et al. 1991, Nicholls & Leatherman 1996, Gornitz et al. 2002). This challenge, in turn, depends on accurate determinations of the effect of accelerated SLR on the natural (physical and ecological) environment. In fact, the National Assessment of Coastal Vulnerability to Future SLR (Thieler & Hammar-Klose 2000) states that determining the physical response of the coast to SLR constitutes "one of the most important problems in applied coastal geology today." Consequently, studies have used various SLR scenarios to explore the socioeconomic, physical, and ecological impacts on coasts in the United States (e.g., Weggel et al. 1989, FEMA 1991, Titus et al. 1991, Yohe et al. 1996) and throughout the world (e.g., Walsh et al. 2004, Nicholls & Tol 2006).

Rising sea level is affecting coastlines throughout the world; the magnitude and types of impacts are related to the geologic setting and physical and ecological processes operating in that environment. Unlike infrequent, large-magnitude storms that can change the complexion of the coast in a few hours (e.g., Mississippi coast due to Hurricane Katrina, **http://coastal.er.usgs.gov/hurricanes/katrina/photo-comparisons/mainmississippi.html**), impacts attributed solely to SLR are usually slow, repetitive, and cumulative. This paper reviews the state of knowledge concerning the response of coasts to SLR, and concentrates on coastal plain settings, including beaches and barrier chains, and associated tidal inlets and backbarrier wetlands.

SEA-LEVEL TRENDS

Sea level is a function of the ocean surface, which is controlled by the (*a*) volume of ocean water, (*b*) volume of the ocean basins, and the (*c*) distribution of the ocean water. Moreover, crustal deformation and sediment compaction control coastal land elevation, which, in turn, controls sea level. Many elements influence sea level. These elements operate globally and locally over a wide range of timescales including days to weeks (tides, storms), seasonal (steric changes, weather), 10°–10^{4} years (climate, tectonic), and up to millions of years (ocean basin evolution). Two primary factors dictate the present rate of SLR: (*a*) thermal expansion due to heat uptake by ocean surface waters and (*b*) water input caused by the transfer of water from the land to the oceans (IPCC 2007). **Supplemental Appendix 1** outlines recent atmospheric and oceanic warming trends (follow the Supplemental Material link from the Annual Reviews home page at **http://www.annualreviews.org**).

Supplemental Material

Globally, sea level has risen approximately 120 m since the last glacial maximum approximately 20,000 years ago (Fairbanks 1989), and it reached a near standstill 2000 to 3000 years ago when the rate of SLR slowed to 0.1 to 0.2 mm year^{-1} (Lambeck & Bard 2000). Global warming has led to thermal expansion of the ocean and a net influx of water from melting glaciers. Tide gauge records and other sea-level proxy indicate that from 1870 to 2004, sea level rose by 195 mm, with an average rate of rise of 1.7 +/− 0.3 mm year^{-1} and an acceleration of 0.013 +/− 0.006 mm year^{-1} (Church & White 2006) (**Figure 2**). Based on 177 tide gauge stations for the 1948 to 2002 period, Holgate & Woodworth (2004) estimated a SLR rate of 1.7 +/− 0.9 mm year^{-1}, giving a cumulative rise of 9.2 cm +/−. The most recent records of sea-level change consist of altimetry data from TOPEX/Poseidon and Jason satellites (Nerem & Mitchum 2001). For the 10-year period between 1993 and 2003, satellite altimetry data show a SLR rate of 3.1 +/− 0.7 mm year (Cazenave & Nerem 2004, Leuliette et al. 2004).

Estimates of the various contributions of the present and past rates of SLR are presented in **Table 1** (IPCC 2007). Thermal expansion accounts for more than half of the present (1993–2003) SLR trend (1.6 +/− 0.5 mm year^{-1}) caused by warming to a depth of 3000 m. The influx of water by melting glaciers is about half that value (0.77 +/− 0.22 mm year^{-1}). Comparatively, lesser amounts of water come from the Greenland and Antarctic ice sheets, which combined, store enough water to raise sea level by 63.9 m (Bamber et al. 2001, Lythe et al. 2001). The IPCC (2007) estimates that sea level will rise from 0.18 to 0.59 m relative to the 1980–1999 position by the end of this century (**Table 2**). This range is based on different atmospheric-ocean global circulation models using various warming scenarios.

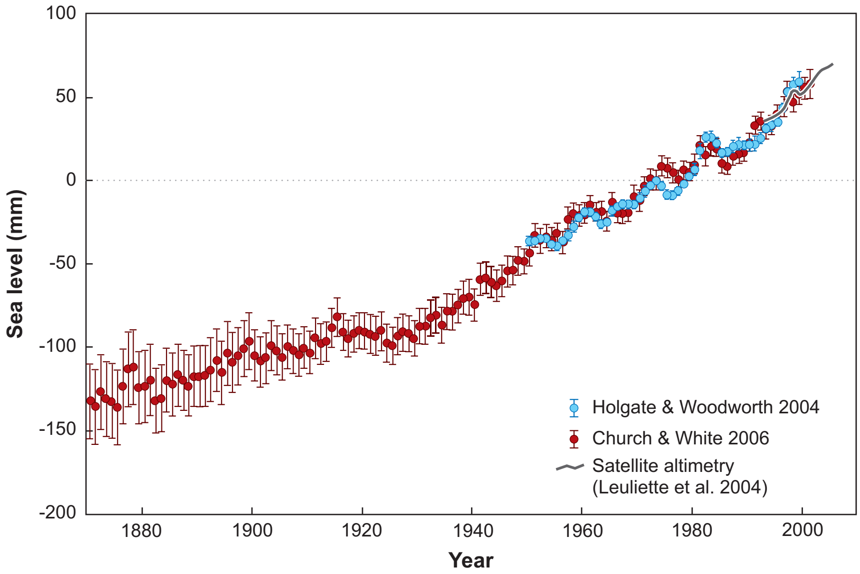


Figure 2

Annual averages of global mean sea level from IPCC (2007). The red data are updated from Church & White (2006); the blue data are from Holgate & Woodworth (2004), and the gray curve is based on satellite altimetry from Leuliette et al. (2004). Error bars show the 90% confidence limits. Zero represents the 1961–1990 averages for red and blue data. Gray curve represents a deviation from red data for the period 1993–2001.

ISLAND AND LOWLAND INUNDATION

Inundation constitutes one of the dramatic and immediate affects of SLR on low-lying coastal areas around the world (Bird 1993). Inundation includes the flooding of deltaic regions, saltwater incursion into regional coastal areas and many urban centers, and disruption of transportation (**Figure 1**; Titus 2002). In the heavily populated state

Table 1 Contributions to SLR (IPCC 2007)

Source of seal level rise	Rate of SLR (mm year^{-1}) 1961–2003	1993–2003
Thermal expansion	0.42 ± 0.12	1.6 ± 0.5
Glaciers and ice caps	0.50 ± 0.18	0.77 ± 0.22
Greenland ice sheet	0.50 ± 0.12	0.21 ± 0.07
Antarctic ice sheet	0.14 ± 0.41	0.21 ± 0.35
Sum of individual climate contributions to SLR	1.1 ± 0.5	2.8 ± 0.7
Observed total SLR	1.8 ± 0.5[a]	3.1 ± 0.7[a]
Difference (observed minus sum of estimated climate contributions)	0.7 ± 0.7	0.3 ± 1.0

[a]Data prior to 1993 are from tide gauges and after 1993 are from satellite altimetry.

Table 2 Projected trends of SLR based on different warming scenarios (IPCC 2007)

Case	Temperature change (°C at 2090–2099 relative to 1980–1999)[a]		SLR (meters at 2090–2099 relative to 1980–1999)
	Best estimate	Likely range	Model-based range excluding future rapid dynamical changes in ice flow
Constant year 2000 concentrations[b]	0.6	0.3–0.9	NA
B1 scenario	1.8	1.1–2.9	0.18–0.38
A1T scenario	2.4	1.4–3.8	0.20–0.45
B2 scenario	2.4	1.4–3.8	0.20–0.43
A1B scenario	2.8	1.7–4.4	0.21–0.48
A2 scenario	3.4	2.0–5.4	0.23–0.51
A1F1 scenario	4.0	2.4–6.4	0.26–0.59

[a]These estimates are assessed from a hierarchy of models that encompass a simple climate model, several Earth models of intermediate complexity (EMICs) and a large number of atmosphere-ocean global circulation models (AOGCMs).
[b]Year 2000 constant composition is derived from AOGCMs only.

of New Jersey, 142 km^2 of wetlands lie below 0.61 m. This vulnerable area increases to more than sixfold if regions lower than 2.90 m in elevation are included (Cooper et al. 2005). These authors also report that more than 200 km^2 of residential and urban areas are below 3.0 m (**Figure 3*a***). As seen in North Carolina (**Figure 3*b***), the slope of the land surface governs the extent of flooding in a regime of SLR. A much greater land area will be flooded behind the Outer Banks of North Carolina by an initial 1.5 m SLR compared with a subsequent SLR of equal magnitude due to the greater slope of the land between 1.5 and 3.0 m in elevation.

An acceleration in SLR will highly impact the densely populated Nile River deltaic. In this region, a SLR of 0.5 m will affect 3.8 million people and 1800 km^2 of croplands and a 1.0 m increase will impact 6.1 million people and 4500 km^2 of fertile lands (**Figure 4**; Simonett & Sestini 2007). In Bangladesh, one of the world's most densely populated countries, a 1.5 m rise in sea level is predicted to affect 22,000 km^2 of deltaic area inhabited by more than 17 million people (WED 2004; **http://www.bdix.net/sdnbd_org/world_env_day/2004/bangladesh/pollutions/rise_sea.htm**).

In the West African country of Cameroon, where 25% of the population (4 million people) live in coastal lowlands (6.5% of Cameroon's total area), SLR will have a profound effect on some of the fastest-growing metropolitan areas and ecologically important coastal wetlands (Asangwe 2007). Saltwater incursion is expected to alter maritime swamps, decrease water quality, and spread diseases such as Malaria. Cameroon and many other developing countries around the world cannot afford reclamation projects on a scale similar to those in the Netherlands or the Gulf of Mexico. As a consequence, retreat from the low elevation coastal zone (LECZ; McGranahan et al. 2007) to the hinterlands may be the only option for many coastal communities.

The effect of SLR on small islands and archipelagos has received increased attention due to vulnerability of their coastal resources and populations to SLR-related coastal flooding (Roy & Connell 1991, Leatherman 1997, Smithers & Woodroffe

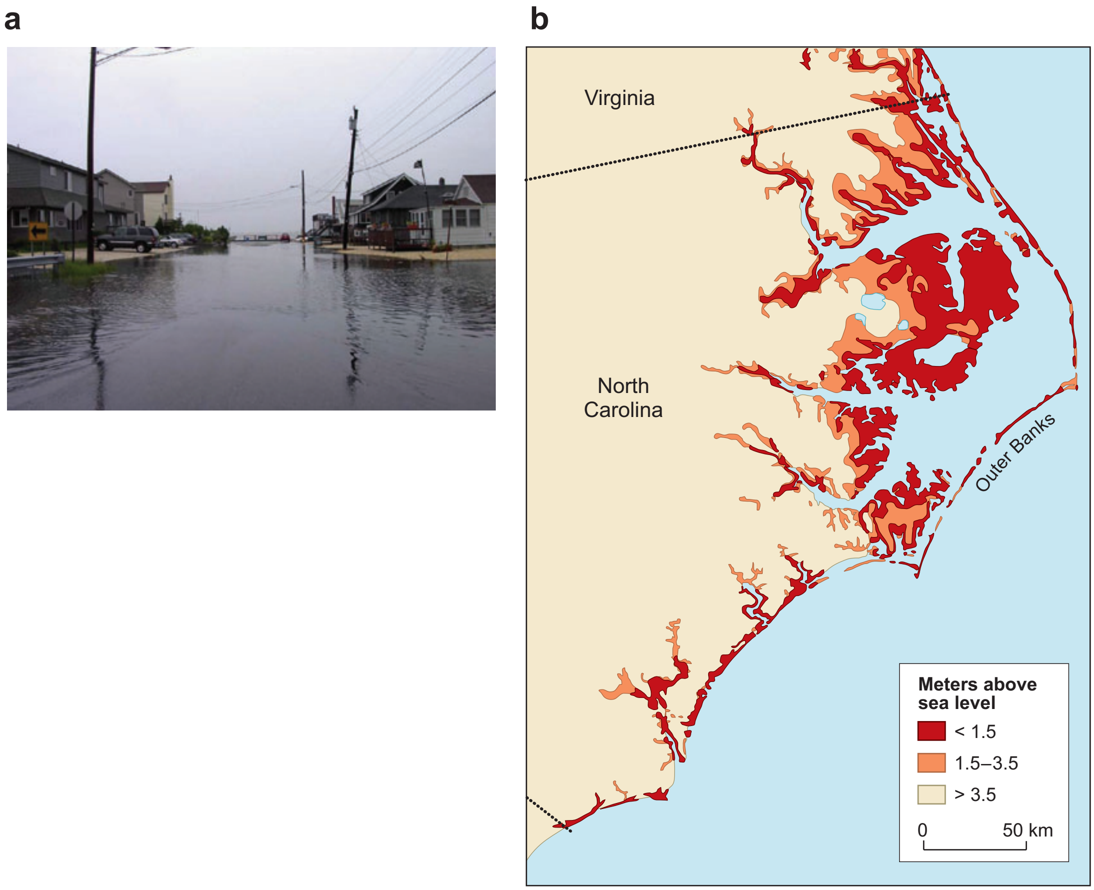


Figure 3

(*a*) Flooded street in Ship Bottom, New Jersey (photo taken at neap tide accompanied by strong winds; from Titus 2002). (*b*) Map of low-lying areas in North Carolina (after Titus & Richman 2001). Note the vulnerable barrier islands of the Outer Banks and wetlands that support ecologically important communities.

2001, Dickinson 2004). Whereas some island nations have coped with SLR and are presently experiencing slightly falling sea level (e.g., Maldives, Mörner et al. 2004; but contended by Woodworth 2005), most of them are affected by global SLR. Because most economic activities on South Pacific islands are concentrated in the coastal zone, including the capitals of many island states (Fiji, Western Samoa, Tonga, etc.), coastal inundation is already driving the patterns of relocation for many residents (Nunn & Mimura 1997). In a recent comprehensive study of Pacific atolls, Dickinson (2004) concludes that SLR may cause overtopping and submergence of relict mid-Holocene paleoreef flats, which would drastically increase wave erosion. This process is an example of the adverse consequences of outer-reef inundation, which would precede the actual flooding of subaerial parts of the islands and atolls.

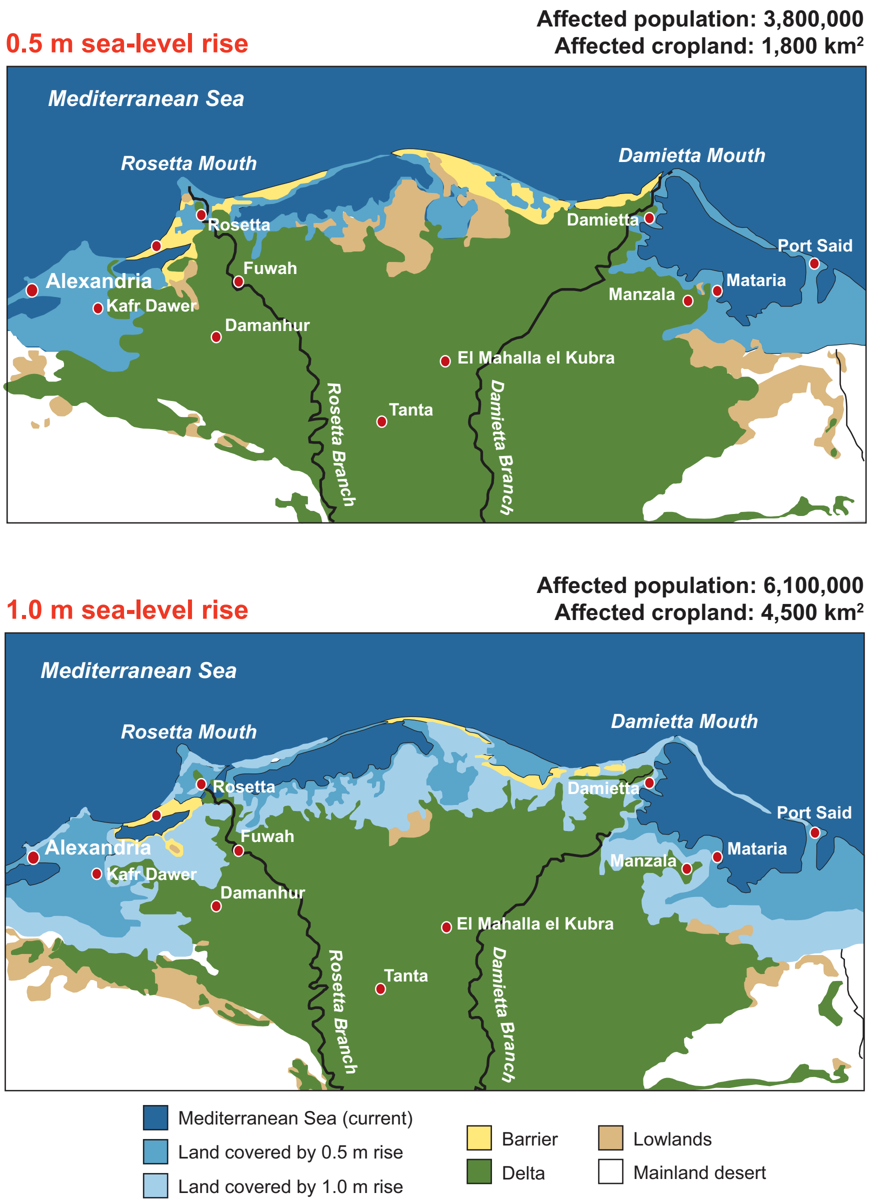


Figure 4

Effect of sea-level rise on the populations and croplands of the Nile Delta (from Simonett & Sestini 2007; **http://maps.grida.no/go/graphic/nile_delta_potential_impact_of_sea_level_rise**).

In addition to inundation, SLR will cause saltwater intrusion into coastal aquifers, particularly in regions of high groundwater withdrawal. For the populations of small islands, reduction or disappearance of potable water may be the greatest impact on their survival, rivaling in importance both coastal erosion and lowland flooding. Entire island nations are already being affected by saltwater intrusion (Tuvalu, Marshall Islands, etc.; Roy & Connell 1991, Nunn & Mimura 1997) and this hazard should be considered together with surface flooding associated with SLR.

SEA-LEVEL RISE AS A DRIVER OF BARRIER ISLAND AND SHORELINE RESPONSE

The various ways that beaches respond to changes in sea level complicate assessments of the impact of SLR on natural systems. Additionally, the perspective of an observed response to SLR, and/or our interpretation of a measured response to SLR, changes as a function of the different timescales. For example, barrier islands can migrate landward for tens to hundreds of kilometers over timescales that span 10^3–10^4 years. Under this scenario, SLR causes the shoreline to move landward during inundation of a stable or subsiding terrestrial land surface if the sediment supply rate cannot keep pace with the SLR rate (Curray 1964). Geologic evidence shows that during the Holocene transgression, coastal barriers, bays, marshes, and wetlands have all experienced dramatic changes that included overstepping and continuous rolling-over. At the other end of the temporal spectrum, the daily physical forces associated with SLR do not appear to contribute to the net coastal sediment transport regime. During decadal- to centennial-scale time periods (10^1–10^2 years), additional processes, such as El Niño, storm surges, tidal inlet processes, seasonal wave climate variations, or human interaction, can either synergistically, episodically, or periodically overwhelm the impact of SLR work to produce beach morphological responses (e.g., Storlazzi & Griggs 2000). Moreover, the global effects of SLR on coasts will vary spatially (Gornitz 1991). As a consequence, some coastal scientists have advocated analyzing and predicting coastal changes on a more local scale (Dean 1987a, Fenster & Dolan 1993, Pilkey et al. 1993, Pilkey & Davis 1987, Cooper & Pilkey 2004). **Supplemental Appendix 2** details the timescales of coastal changes in general and beach and nearshore profiles in particular.

Supplemental Material

The need to predict and manage the potential impact of SLR on coasts necessitates accurate models. Statistical modeling usually involves projecting historical shoreline changes into the future (e.g., NRC 1990, Fenster et al. 1993, Douglas et al. 1998). The National Research Council's (1990) synthesis on managing coastal erosion refers to this approach as historical trend analysis. This one-dimensional, response-based approach uses time as a surrogate for processes, and coastal response is measured by trend delineation of historical shoreline positions (Dolan et al. 1991). Despite this limitation, Leatherman (1984) suggested using historical trends, coupled with an estimate of SLR, to develop a simplified mathematical rule capable of predicting shoreline recession rates. This approach assumes that SLR entirely controls shoreline trends. Leatherman et al. (2000) and Zhang et al. (2004) expanded the concept to include a direct relationship between SLR and shoreline recession over centennial to decadal timescales, but these ideas have met resistance in the coastal community (Pilkey et al. 2000, Sallenger et al. 2000, Cooper & Pilkey 2004). Conceptual, empirical, and mathematical models are also used extensively to predict the impact of SLR on coasts. Although these models attempt to link responses with the processes responsible for those changes, each has limitations. In general, the coastal sciences do not have a holistic model available to make those links reliably in multiple settings (e.g., Cooper & Pilkey 2004, Cowell et al. 2006, Fenster 2006, Stolper et al. 2005). Given this constraint, we present several of the most widely used models to link sea-level changes to coastal responses in general and shoreline migration in particular and the fundamental concepts upon which these models rely.

Equilibrium Profile: A Central Concept

Most two- and three-dimensional models that predict the response of beaches and barriers to SLR in general and predict shoreline migration in particular depend on two primary assumptions: (*a*) beaches and barriers operate as closed material transfer systems and (*b*) the combined beach and offshore two-dimensional profile maintains a constant or equilibrium shape over the long term (i.e., greater than those changes that occur seasonally). These two conditions enable the models to account for all movement of material within an area circumscribed by longshore and offshore boundaries and constrain erosion and deposition to a landward or shoreward direction (Bruun 1988). In essence, these conditions stipulate that a longshore and cross-shore equilibrium exists in which fluctuations to the profile produced by the wave climate (i.e., seasonal variations or storms) and currents maintain the equilibrium profile at a given water-level position, but do not alter its average form over the long-term ($\geq 10^1$ year).

The equilibrium profile has remained a central component of models that predict shoreline retreat based on SLR since the inception of the Bruun rule (Bruun 1954, 1962). The rule is also one of the central themes in coastal geology and sedimentation (Swift & Palmer 1978). Fenneman (1902) and Johnson (1919) first speculated that shore profiles along unconsolidated coasts develop a constant, concave upward shape following adjustments of current, slope, and sediment load. The equilibrium profile is a time-averaged response to variations in energy and sediment supply and consists of a concave upward, exponential shape with a shorter and steeper limb near the shoreline and longer, flatter limb farther offshore (e.g., Moore & Curray 1964, Dean 1991, Larson 1991). Coastal scientists have distinguished between an ideal profile and a dynamic equilibrium profile (NRC 1987, Pilkey et al. 1993) where the former represents a true geometric equilibrium profile shape, and the latter undergoes time-dependent profile adjustments about a long-term average shape. In any case, more than a century of research has shown that coastal systems tend toward equilibrium configurations with defined geometries, over decadal and greater time periods, despite the inherent variability of seasonal and/or episodic nearshore processes (Fagherazzi et al. 2003).

Using curve-fitting techniques through bathymetric data from the Denmark and southern California coasts, Bruun (1954) first described the geometry of the equilibrium profile with a power function equation that relates profile shape at a water depth,

h, to distance offshore (x) and a scale factor, A, constrained primarily by sediment characteristics:

$$h(x) = Ax^{2/3}. \tag{1}$$

Dean (1977) verified this relationship using a profile data set from Hayden et al. (1975) and corroborated the exponent value of 2/3 originally provided by Bruun (1962). Moreover, Dean (1977) found that the shape factor, A, depends primarily on sediment characteristics and speculated that the monotonic profile form is consistent with uniform wave energy dissipation per unit volume of the water column within the surf zone.

The concept of equilibrium profile has not met complete acceptance in the coastal sciences. Arguments against the concept include (*a*) the inability to differentiate between relict and active continental shelves (Dietz 1963); (*b*) the notion that sediment characteristics (e.g., scale factor A in Equation 1), including grain size and fall velocity, can determine profile shape is unrealistic outside the surf zone (Bruun 1988, Pilkey et al. 1993, Thieler et al. 2000, Pilkey & Cooper 2004, Cooper & Pilkey 2004); (*c*) determination of the parameters—including grain size—comes from a nonphysics-based, empirical approach that has omitted germane results (Pilkey et al. 1993, Thieler et al. 2000, Pilkey & Cooper 2004, Cooper & Pilkey 2004); (*d*) limitations of mathematical models that predict shoreline retreat as a function on SLR—e.g., nonsandy antecedent geologic conditions, cross-shore variations in sediment type, sediment-starved environments, offshore sediment losses beyond the active sediment transport prism or nearshore, tectonically or isostatically active coasts, a model's inability to provide the kinematics and/or dynamics required to maintain the equilibrium profile (Pilkey et al. 1993, Thieler et al. 2000, Pilkey & Cooper 2004); and (*e*) the use of two-dimensional models in a three-dimensional system and the problems associated with system boundaries. **Supplemental Appendix 3** summarizes the laboratory, empirical, and modeling studies used to confirm the existence and quantify the shape of the equilibrium profile. **Supplemental Appendix 4** discusses issues related to longshore and cross-shore boundary conditions and presents modifications to the equilibrium beach profile that have attempted to improve its capability of predicting the response of barriers and shorelines to SLR.

Supplemental Material

Predictive Models of the Sea-Level Rise on Beaches and Barriers

Coastal scientists, engineers, planners, and managers use both two- and three-dimensional models to determine the impact of SLR on beaches and barriers. The two-dimensional shoreline response models include the Bruun rule (Bruun 1962) and modifications to the Bruun rule (Dean & Maurmeyer 1983, Dubois 1992, Bray & Hooke 1997, Davidson-Arnott 2005). Komar et al. (1991) differentiate between the Bruun rule (as proposed by Schwartz 1967) and the Bruun model by referring to the former as the driving equations that predict shoreline retreat and the latter as the model's assumptions. The Bruun rule is given as

$$R = \frac{L_*}{B + h_*} S = \frac{1}{\tan\theta} S, \tag{2}$$

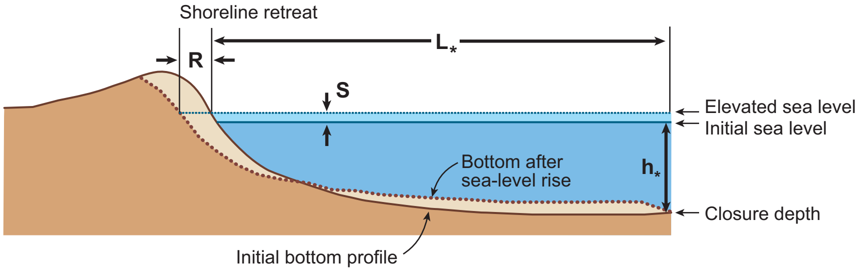


Figure 5

The Bruun rule of shoreline retreat (after Cooper & Pilkey 2004).

where the shoreline retreat rate is a function of the cross-shore distance L_* to closure depth h_*, the berm height or elevation estimate of the eroded area, B, and the SLR rate, S (**Figure 5**). Consequently, $L_*/B + h_*$ reduces to $1/\tan\theta$. As a result, small values of $\tan\theta$, typical of many coastal regions (e.g., 0.01–0.02), produce an $R = 50S$ to $100S$ as a rule of thumb. This generalization suggests that small increases in SLR result in relatively large shoreline recessions. The Bruun model assumes an upward and landward translation of an equilibrium profile, conservation of mass applies across the shore (i.e., between the upper profile and lower shoreface) and along the shore (i.e., no sediment budget adjustments), and that closure depth exists. Thus, as the profile shifts landward, eroded upper beach sediment is transported offshore and deposited such that the vertical accretion equals the magnitude of SLR.

Because closure depth, h_* in the Bruun rule (and modifications to the Bruun rule), determines the value of L_*, the cross-shore distance to h_*, uncertainty in shoreline position predictions will vary systematically. For example, overestimates of h_* will result in longer cross-shore profiles, L_*, and consequently produce larger values of shoreline recession (Komar et al. 1991). However, as mentioned above, the Bruun rule is not applicable when the nearshore receives sediment from outside the nearshore or closure depth (Dean 1987b). Determining the processes and responses involved in profile translation at individual sites, including the offshore extent (i.e., distance and flux) to which sediment transport and deposition occurs, and the time frames over which these phenomena occur can minimize shoreline position prediction uncertainty.

Modifications of the Bruun rule have attempted to attain greater accuracy in representing the beach profile's response, or zone of active sediment transport, to SLR. For example, Dubois (1977) incorporated more complexity into the cross-shore subaerial and subaqueous beach profile (i.e., adding an offshore bar). Dean & Maurmeyer (1983) and Dean (1991) generalized the Bruun rule by expanding the spatial scope of the profile to include the entire barrier island system:

$$R = \frac{L_{*_o} + W + L_{*_L}}{(h_{*_o} - h_{*_L})} S, \qquad (3)$$

where L_{*_o} and L_{*_L} are the ocean and lagoon active nearshore widths, respectively; W is the barrier island width; and h_{*_o} and h_{*_L} are closure depths on the ocean and

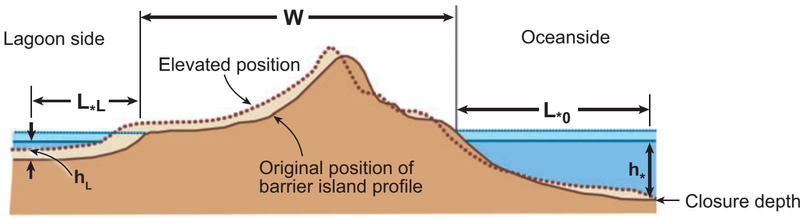


Figure 6

The Generalized Bruun rule of Dean & Maurmeyer (1983) showing the entire barrier island system maintaining its form while migrating landward under the influence of sea-level rise.

lagoon sides of the barrier (**Figure 6**). The equation reduces to the Bruun rule if no deposition occurs on the barrier or in the lagoon.

Komar et al. (1991) suggested that a three-dimensional sediment budget can improve shoreline retreat rate predictions as a function of SLR. The sediment budget approach calculates sediment sources and sinks within a control volume and, therefore, accounts for the longshore transport gradient. For example, Hands (1983) developed a model based on the Bruun rule that contains an overfill ratio[1] for fine-grained sediment transport, F_A. If $F_A > 1$, more erosion, R, is needed to account for fine-grained sediments that are transported away from the active profile:

$$R = \frac{L_* F_A}{(B + h_*)} S + \frac{\sum Q_s}{Y(B + h_*)}, \tag{4}$$

where $\sum Q_s$ provides a longshore sediment discharge gradient of length, Y, into and out of a control volume along a beach with berm height, B. Komar et al. (1991) include a term (G_B) to account for the longshore sediment transport gradient $\partial Q_s/\partial Y$. When $\partial Q_s/\partial Y = 0$ or constant, the three-dimensional models reduce to two dimensions. However, empirical evidence does not ubiquitously support this mathematical statement of equality (Fenster & Dolan 1993). Everts (1985) used the term V_0 to account for changes to the volume of sand (and, consequently, to the shoreline position) within the control volume:

$$V_0 = \Delta t \sum_{i=1}^{n} \frac{\Delta V_i}{\Delta t}, \tag{5}$$

where V_i is the volume of sand entering or leaving the control volume over time, t, and Δt is a result of n transport processes. This term seeks to consider all longshore and cross-shore processes that influence profile and shoreline changes, including tidal inlet controls, overwash, and anthropogenic factors. This approach provides an additional term to account for all profile changes not attributable to SLR so that the sediment balance can determine volume changes attributable to SLR, V_l:

$$kV_l + V_O - (V_g + V_g') = 0, \tag{6}$$

[1]The term overfill ratio used by Hands (1983) differs from the term used by the NRC which "quantifies the amount of material to be placed on the beach to yield a unit volume of compatible beach sand. This factor allows the composition of the eroding beach or bluff material to be included."

where k is the volume of sand-sized or larger-grained sediment in V_l, and V_g and V_g' are the volumes of sand-sized material on the backbarrier and shoreface, respectively, and are derived from V_l. Despite this modeling approach, the NRC (1987) claims that the lack of reliable data, such as annualized values of longshore transport volumes, tidal inlet affects, overwash, and offshore volumetric losses/gains, can constrain these approaches.

Bruun (1988) presents a discussion of the two- and three-dimensional uses of the Bruun rule. Komar et al. (1991) provide a more comprehensive evaluation of the two- and three-dimensional beach response to SLR models that existed prior to 1991. Cooper & Pilkey (2004) provide a comprehensive review of the Bruun rule. Davidson-Arnott (2005) present the most recent revision of shoreline retreat models as a function of SLR.

Predictive Morphological-Behavior Models (Large-Scale Coastal Behavior Models)

Beginning in the 1990s, a suite of quantitative large-scale coastal behavior (LSCB) models were developed that aimed to simulate the large-scale morphologic and stratigraphic evolution of coasts that results from changes in sea level and sediment supply (Cowell et al. 1995; Niedoroda et al. 1995; Stive & de Vriend 1995; Stolper et al. 2005; Moore et al. 2006, 2007). Similar to the shoreline response models that use time as a surrogate for processes, the LSCB models utilize geometric cross-shore profile parameters as a proxy for processes. Specifically, these conservation of mass–shoreface translation models are governed by an SLR scenario and use a set of parameters that rely on the equilibrium slope concept, an initial volume of (shoreface, barrier, and estuary) sediment, a substrate slope, and a substrate sand content. The original shoreface translation model (STM) (Cowell & Thom 1994, 1995) sought to deliver numerical solutions for profile kinematics while simulating the effects of geological inheritance—one of the criticisms of the Bruun rule—as well as storm processes, sea-level changes, and variable sediment budgets. The goal of the model is to minimize the number of model parameters that govern large-scale coastal evolution (to reduce uncertainty), but to maximize the model's potential to capture system complexity. Spatially, the models can provide site-specific analysis at one profile, large-scale analysis because the model aggregates the spatial variability of an entire coastal cell into one shore-normal profile (that presumably captures both boundary conditions and driving forces), and a quasi-three-dimensional application.

Stolper et al. (2005) sought to improve on the prototype STM of Cowell & Thom (1994) and Cowell et al. (1995) with the GEOMBEST model (Geomorphic Model of Barrier, Estuarine, and Shoreface Translations). This model defines the substrate by stratigraphic and sedimentologic properties such as erodibility (graded resistance to the erosion potential) and composition instead of requiring an unlimited unconsolidated sediment supply. This quality enables the model to assess the geological framework when simulating morphological evolution and shoreline migration. Unlike models that depend on equilibrium profiles (as explained above), GEOMBEST can account for the disequilibrium found on some shorefaces and, in fact, is driven by the disequilibrium produced sea-level changes and the vertical displacement of the profile (e.g., Pilkey et al. 1993; Wright 1995; Thieler et al. 1995, 2000). Validation of these models, and a simulated evolutionary history of a particular setting for a given sea-level trend, comes through inverse modeling, whereby a known stratigraphy and residual surface serve as the end member toward which trial and error simulation experiments move.

Moore et al. (2006, 2007) used GEOMBEST to determine that SLR was the single greatest causative factor in controlling the evolutionary history of the Outer Banks near Cape Hatteras followed by changes in the sediment budget. Additionally, Moore et al. (2007) used GEOMBEST to predict future impacts of various SLR scenarios for AD2100 on LSCB. Model runs using low, middle, and high IPCC (2001) SLR estimates showed that a 0.9 m increase in sea level will increase erosion rates of up to 2.5 times the present-day rates found in erosion hot spots (Morton et al. 2005). However, Moore et al. (2007) conjectured that the barrier could remain intact given that rapid migration currently exists on other barriers (Morton et al. 2004; Penland et al. 2005). However, simulation runs using the 1.4 m–1.9 m rise in sea level by AD2100 show that threshold collapse and subsequent drowning could occur along the Outer Banks similar to the modern day analog of Louisiana's Isle Dernieres.

WETLANDS

Background

Coastal wetlands have increasingly been recognized as a unique and vulnerable habitat as they represent a transitional zone between tidal flats and uplands that is exposed to extremes of temperature, salinity, and tidal inundation. Salt marshes cover extensive areas of estuarine and deltaic environments in mid to high latitudes and are threatened worldwide by accelerating SLR (Church & White 2006).

Delicate balances between accretion and subsidence, bioproductivity and decomposition, erosion and vegetative stabilization, and tidal prism and drainage efficiency maintain tidal marshes (Cahoon et al. 1995, Morris et al. 2002, French 2006). These processes are in turn controlled by complex interrelated feedbacks with physical parameters including climate, sea level, and regional tectonics (**Figure 7**). Salt marsh morphology is a function of, and a control on, these parameters.

Salt marsh morphology can generally be classified as either ramped or platform, **Figure 8**, corresponding to a predominance of tidal sedimentation or bioproductivity, respectively, as the key driver of accretion (Allen 2000, French 2006). Marshes dominated by supratidal marsh exhibit a platform-like morphology (**Figure 8*a***). Supratidal marsh peat is more highly organic than intertidal marsh peat, reflecting lower influx of inorganic sedimentation due to reduced frequency, duration, and depth of tidal inundation (Argow 2006). Intertidal marshes, dominated by low marsh, may exhibit ramped (**Figure 8*b***), platform, or mixed morphology, depending on the relative influx of inorganic sedimentation, tidal range, and wave climate (Allen 2000, Morris et al. 2002).

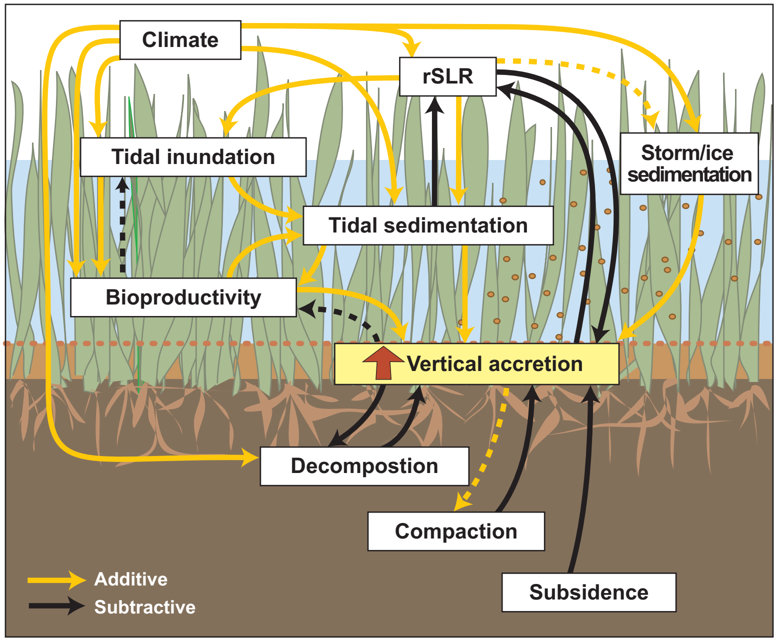


Figure 7

Conceptual model of the major factors affecting marsh elevation (after Argow 2006).

Vertical Accretion

In order to understand and predict the impact that rising sea level may have on salt marshes, it is necessary to quantify the processes controlling the elevation and evolution of the marsh surface. Vertical accretion may be defined as net vertical growth of the marsh surface relative to a stationary datum, resulting from both mineral sediment influx and the production of organic matter (**Figure 7**).

Tidal sedimentation is a key parameter in this complex system. Controls on inorganic suspended sediment transport and deposition include tidal range and depth of inundation, vegetation density, and rates of particle settling (Allen 2000, Day et al. 1999, French 2006). On short timescales, tidal sedimentation is controlled by tidal currents and the suspended sediment load, as sediment is deposited during high slack water. Over longer timescales, episodic storms may deliver sediment volumes in a single event that are orders of magnitude greater than single-tide sedimentation rates (Leonard & Reed 2002, Donnelly et al. 2004, Cahoon 2006). Marsh accretion over millennial scales is positive during a regime of gradually rising sea level; however, over tidal and seasonal timescales, mineral and organic material on the marsh surface and in tidal channels may be resuspended by rainfall (Wolters et al. 2005), wind-generated waves (Moller 2006), ice effects (Dionne 1989, Argow 2006), storms (Leonard et al. 1995, van de Plassche et al. 2006), tidal currents (French 2006), and modified by the effects of biota (Leonard & Reed 2002).

Suspended sediment deposition is enhanced by salt marsh vegetation, which alters the hydrodynamics of intertidal systems, raising the elevation of the marsh platform

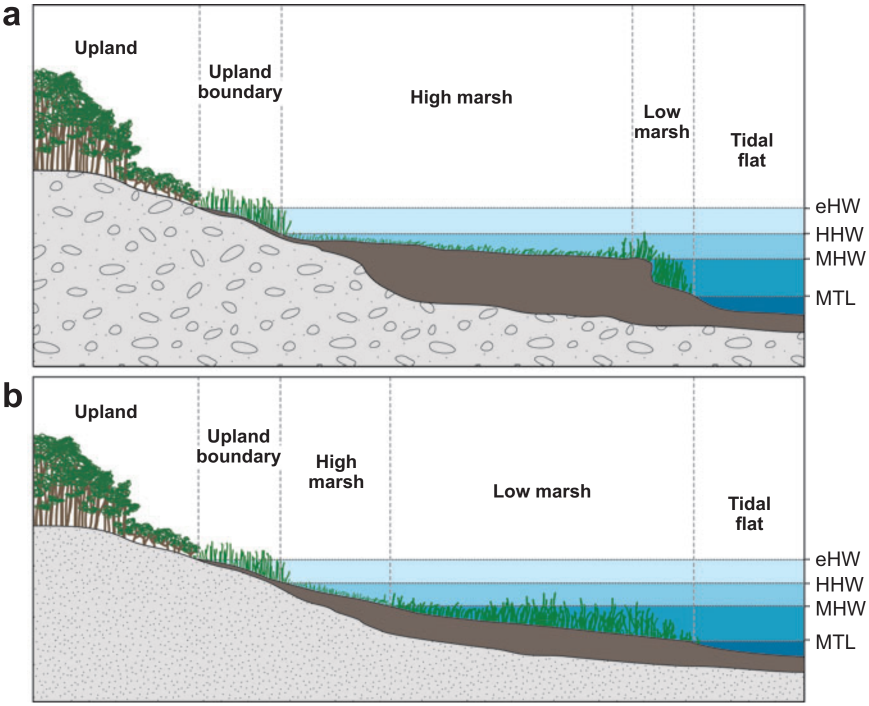


Figure 8

Platform (*a*) versus ramped (*b*) morphologies in tidal wetlands (after Argow 2006). Marsh zones are shown relative to mean tide level (MTL), mean high water (MHW), mean spring high water (HHW), and extreme high water (eHW).

(Nepf et al. 1997, Leonard & Reed 2002). Reduced depth of inundation, in turn, acts as a feedback on both tidal sedimentation and bioproductivity, driving coastal wetlands toward an equilibrium elevation relative to mean sea level (Boorman et al. 2001, Mendelssohn & Morris 2000, Morris et al. 2002, Morris 2007). The sensitivity of marshes to longer-term (decadal to millennial) sea-level variation has long been exploited as a record of paleo–sea level (e.g., van de Plassche et al. 2006); however, short-term marsh response to variable sea level may also be significant (Morris et al. 1990).

The dominant species in the marsh, together with nutrient availability, control the rate of organic peat production. For example, intertidal marsh species (*Spartina alterniflora*) are more productive than high-marsh species (dominantly *S. patens* or *Distichlis spicata*) (Donnelly & Bertness 2001), contributing to higher rates of vertical accretion on intertidal marshes (~6–8 mm year^{-1}) than on high marshes (~2–3 mm year^{-1}) (**Figure 9**; FitzGerald et al. 2006). The highest rates of marsh vertical accretion are found in fluvially dominated systems due to high inorganic sediment influx.

Sedimentation rates, either measured in situ on relatively short (tidal to annual) timescales or approximated from long-term deposition in cores, are used to predict

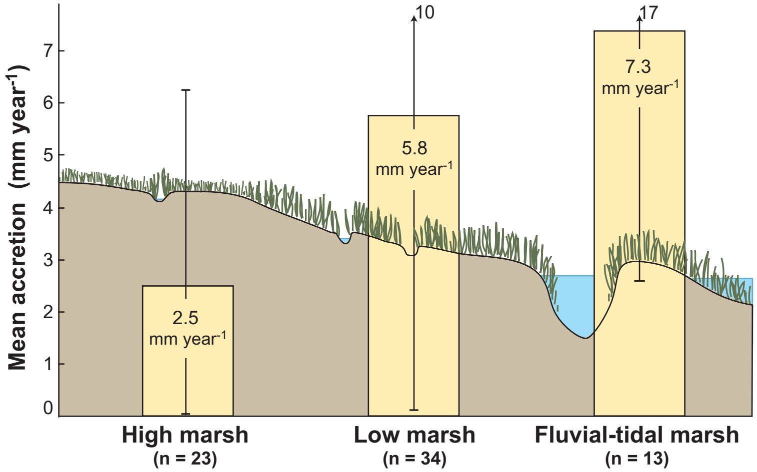


Figure 9

Rates of vertical accretion on tidal marshes based on a survey of the published literature, subdivided by type of marsh. Yellow bars represent mean vertical accretion; whiskers represent the range of reported data. Means ± standard deviation are as follows: high marsh, 2.5 ± 1.4 mm year^{-1}; low marsh, 5.8 ± 2.8 mm year^{-1}; fluvially influenced marsh, 7.3 ± 3.2 mm year^{-1}. After Argow 2006.

numerically future marsh evolution in response to rising sea level despite uncertainties in proxies used or variability in directly measured empirical datasets (e.g., Woolnough et al. 1995, Callaway et al. 1996, Day et al. 1999, Temmerman et al. 2003, Mudd et al. 2004, French 2006, Kirwan & Murray 2007). Supratidal marshes face greater risk of inundation, as average maximum accretion rates are comparable to modern SLR (**Figure 3**). However, these marshes may be rapidly colonized by low marsh vegetation, preserving some functions of the marsh, while reducing biodiversity (Donnelly & Bertness 2001).

Predictive Models of Marsh Accretion

An acceleration in the rate of SLR to 5.0 mm year^{-1} or greater (IPCC 2007) may severely impact wetlands and tidal flats behind barrier island chains, in estuaries, and on lower delta plains. If these environments are unable to accrete vertically at the same rate as rising sea level, then they will be converted to lower intertidal and open water areas. Some of the loss in areal extent will be compensated for by landward migration of these environments, unless upland slope or human infrastructure prevent this (Donnelly & Bertness 2001). The timing and pattern of this evolution has been investigated using numerical models.

Salt marsh vertical accretion is a complex response to multiple factors, and observed spatial and temporal variation remains difficult to capture in existing numerical models (Nyman et al. 1995, Woolnough et al. 1995, Day et al. 1999). Several models discussed below provide insight into the drivers of marsh evolution.

Krone (1987) presented a method for simulating historic marsh elevations, based on elevations of water and marsh surface, suspended sediment concentration, and median settling velocity. This sedimentary infilling approach was later refined for application to other marshes, particularly those with high organic accretion through a series of more complex models. In particular, Morris et al. (2002) developed a model driven not by inorganic sedimentation, but by changes in bioproductivity with varying levels

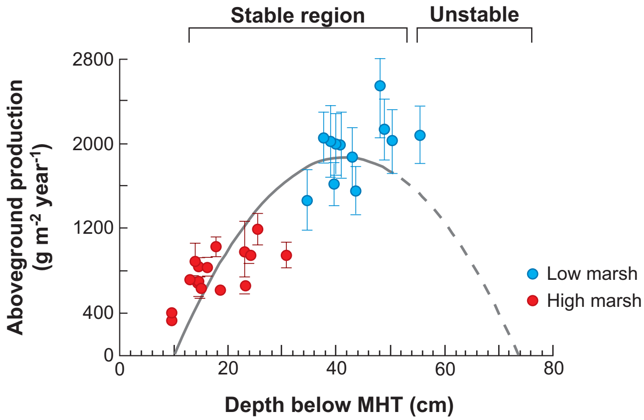


Figure 10

Observed productivity of the low-marsh grass *Spartina alterniflora* versus depth of inundation below mean high tide (MHT) of high-marsh (*red circles*) and low-marsh (*blue circles*) sites. Depth below MHT is a highly significant predictor of productivity ($r^2 = 0.81$, $P < 0.0001$). The solid line represents stable range of equilibrium productivity; the dashed line represents instability and reduced productivity. Local MHW is ~60 cm above MSL. From Morris et al. (2002).

of tidal inundation. Two decades of in situ primary productivity measurements were compared with interannual variation in mean sea level measured by tide gauge to develop a model of optimal inundation for maximum bioproductivity (**Figure 10**; Morris et al. 2002), which in turn was used to drive a model of marsh surface elevation change.

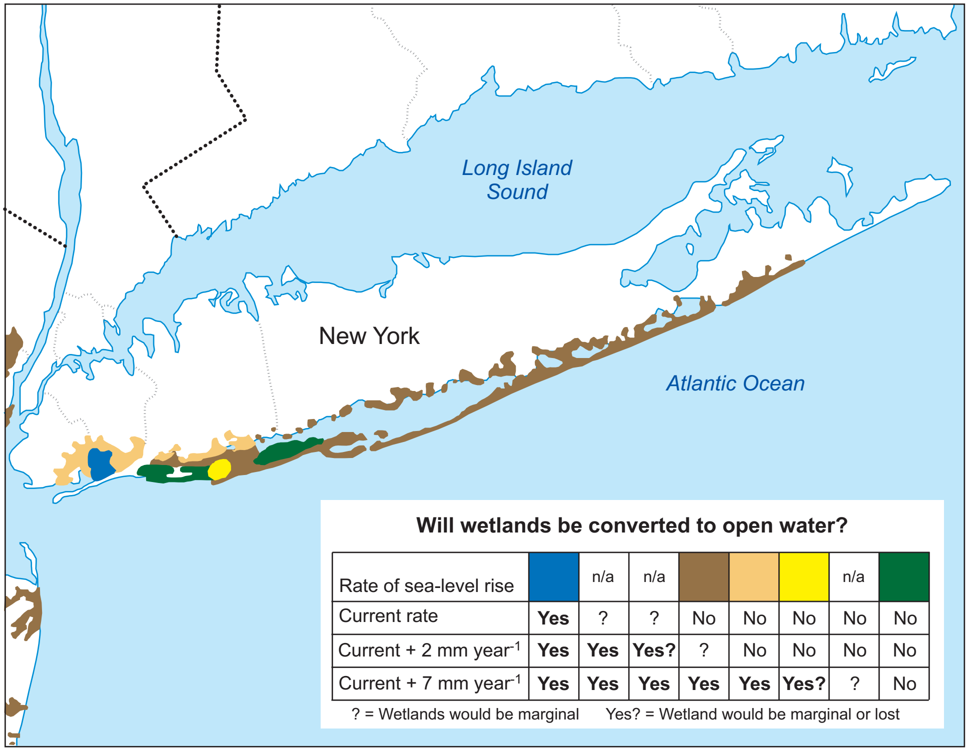
Rybczyk & Cahoon (2002) employ a cohort modeling approach combined with field data to explore the submergence potential of stable and failing Gulf Coast marshes. A primary productivity submodel is integrated with a complex sediment dynamics submodel forced by mineral sediment influx. Changes in rates of decomposition, root distribution, sediment compaction, peat characteristics, and marsh surface elevation were also incorporated. Long-term predictions based on the model suggest that the stable Gulf marsh site will not be able to keep up with accelerating rates of SLR (Rybczyk & Cahoon 2002).

Many models are empirically driven and parameterized based on long-term records from peat cores; however, marsh vertical accretion rates do not necessarily coincide with SLR (French 1994). To counter this problem, French (2006) proposes a zero-dimensional mass-balance model based on an extensive data set from published literature, including rates of SLR, tidal range, and sediment supply. The results show that marshes dominated by inorganic sediment supply are generally near equilibrium with present rates of SLR and are more resilient to increases in the rate of SLR as tidal range and sediment supply increase compared with marshes with little inorganic influx (French 2006).

In contrast, a simplified three-dimensional model from Kirwan & Murray (2007) demonstrates that the presence of vegetation stabilizes marsh surface elevation relative to SLR, suggesting that rates of bioproductivity drive elevation gains sufficient to keep up with SLR. The model assumes a high rate of bioproductivity, but is robust within the constraints of the chosen parameters (Kirwan & Murray 2007). Contradictory model results reflect the complexity of the marsh system and the challenge of generalizing between marsh environments (organic versus sediment-rich).

Predictive Models of Marsh Inundation with Sea-Level Rise

Predictive models are being applied to policy and management decisions along the U.S. mid-Atlantic coast. A field-based predictive modeling study by Reed et al. (2006) utilized the wealth of published vertical accretion data in concert with tide gauge records to predict the likelihood of marsh submergence for three possible SLR scenarios for coastal wetlands extending from Virginia to New York. The results of Reed et al. (2006) suggest that the majority of mid-Atlantic marshes are stable relative to present rates of SLR, and will accrete vertically at a rate comparable to accelerated rates of SLR (**Figure 11**). However, the data used in the analysis are measurements from predominately low marsh environments, thereby possibly overestimating whole-marsh accretion. Even using relatively high accretion rates, high-risk areas are identified wherein the marsh environment is extremely vulnerable to the possible



Rate of sea-level rise		n/a	n/a				n/a	
Current rate	**Yes**	?	?	No	No	No	No	No
Current + 2 mm year^{-1}	**Yes**	**Yes**	**Yes?**	?	No	No	No	No
Current + 7 mm year^{-1}	**Yes**	**Yes**	**Yes**	**Yes**	**Yes**	**Yes?**	?	No

? = Wetlands would be marginal Yes? = Wetland would be marginal or lost

Figure 11

Map of the likelihood of the conversion of wetlands to open water for Long Island given three rates of relative SLR. Lower rates are at the left. From Reed et al. (2006).

acceleration in SLR, and this approach represents a valuable step toward using scientific predictive capabilities to guide site-based management efforts.

In a regime of accelerated SLR, the conversion of marshland to open water is not expected to be linearly related to the rate of SLR. For a given region, a threshold may exist in which a marsh can no longer keep pace with SLR. Once this threshold is exceeded, marshes will succumb to increasing tidal inundation (FitzGerald et al. 2006, Reed 2002). As sections of marsh are converted to open water, local wind-generated waves will erode marsh banks and the rate of marsh loss will hasten (Moller 2006, Wilson 2006). Wave erosion is greatest along shorelines located downwind of the longest local fetch and exposed to the dominant winds. The baffling role of vegetation is important in protecting the substrate, as unvegetated portions of the marsh are susceptible to erosion (Cahoon & Reed 1995). Mature marsh systems in largely infilled backbarriers will experience less erosion than marshes fronting open expanses of water. Inundation and increased erosion of salt marsh peat may provide an additional positive feedback mechanism for increasing greenhouse gases and global warming by converting organic matter to carbon dioxide and methane (Chmura et al. 2003). Marsh loss may precipitate hydrodynamic and sedimentological change affecting an entire coastal system.

The process whereby marshes are transformed to an open water environment is highly complex, site-specific, and related to a number of variables, including suspended sediment concentrations, nutrient abundance, biomass production, storm frequency and intensity, and other factors (Reed 2002). However, gross marsh morphology will influence the temporal and spatial pattern of marsh inundation (Argow & FitzGerald 2006). Marshes exhibiting a platform-like morphology, such as the high-marsh-dominated coastal wetlands of New England, should lose little areal extent during the initial stages of increased rates of SLR. When mean high water levels eventually exceed the elevation of the marsh platform, however, the marsh may rapidly become inundated. Ramped marshes, in contrast, would be expected to show gradual, persistent loss of areal extent (or inland migration) in concert with rising sea level. Barataria Bay, Louisiana, is an excellent example of gradual marsh loss due to the region's high rate of relative SLR and significant wetland loss during the past half century (Barras et al. 1994) (details discussed below).

BARRIERS AND TIDAL INLETS

Barriers comprise approximately 15% of the world's coastlines and are found along every continent except Antarctica, in every type of geological setting, and in every type of climate (Davies 1980, Reinson 1984, Davis 1994). They are best developed in areas of microtidal to mesotidal range[2] and in mid to lower latitudes (Hayes 1979). Climatic conditions control the vegetation on the barriers and in backbarrier regions, sediment type, and in some regions such as the Arctic, the formation and modification of barriers themselves. Generalized barrier environments and stratigraphy are given in **Figure 12**.

[2]Microtidal is a tidal range less than 2 m. Mesotidal is a range between 2 and 4 m (Davies 1980).

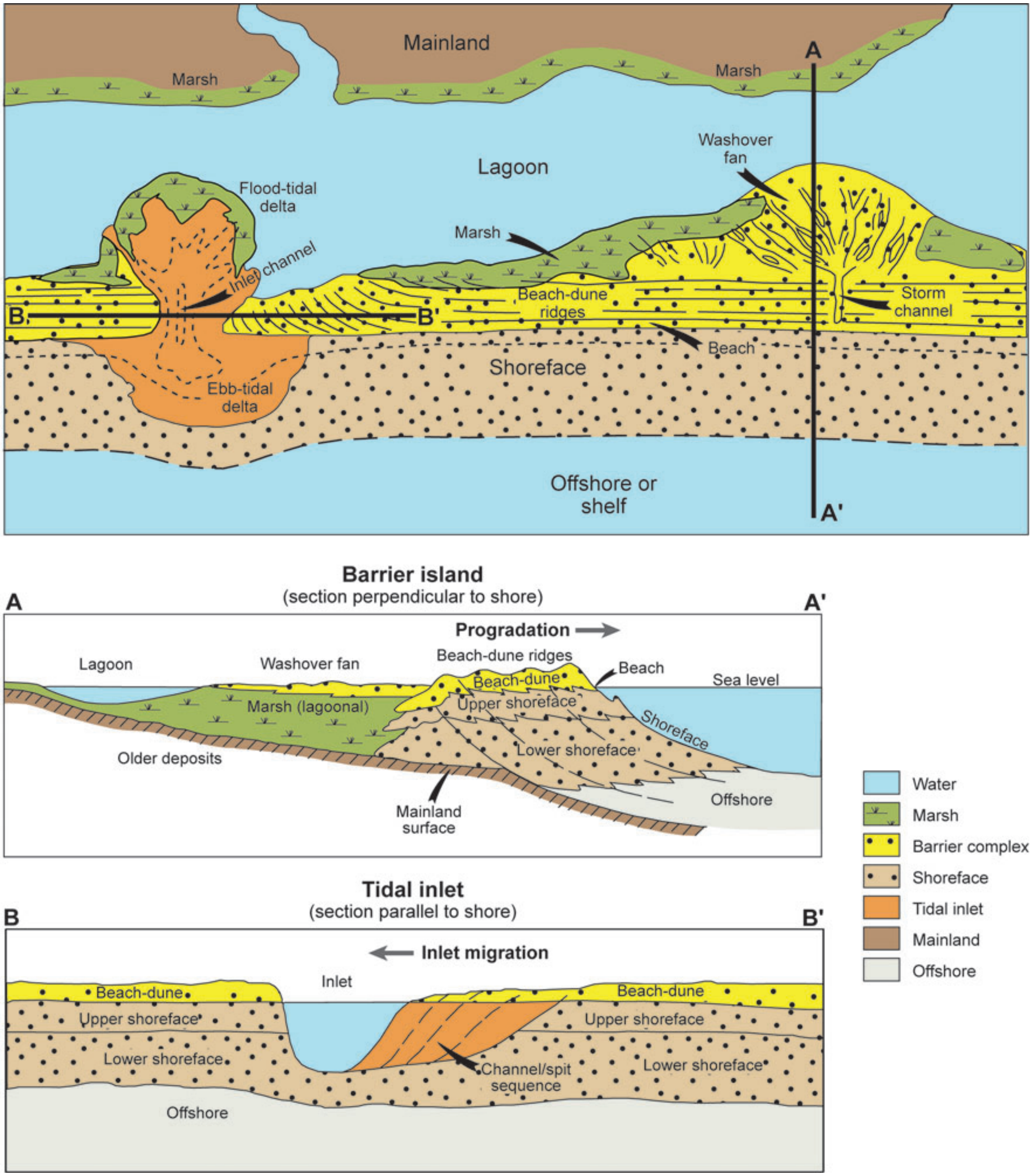


Figure 12

Barrier island and tidal inlet environments (after Pettijohn et al. 1988).

Barriers are linear features that tend to parallel the coast, generally occurring in groups or chains. The longest barrier chains in the world coincide with Amero-trailing edges and include the East Coast of the United States (3100 km) and the Gulf of Mexico coast (1600 km). Sizable barrier chains also exist along the East Coast of South America (960 km), the East Coast of India (680 km), the North Sea coast of Europe (560 km), Eastern Siberia (300 km), and the North Slope of Alaska (900 km). Isolated barriers are common along glaciated coasts, such as in northern New England and eastern Canada, and along high-relief collision coasts, such as the west coast of North and South America. Hayes (1979) classified barrier coastlines based on wave energy and tidal range: (*a*) Wave-dominated coasts contain long linear barrier islands, few tidal inlets, and open-water lagoons and bays (i.e., Texas, the panhandle of Florida, the outer banks of North Carolina, northern New Jersey, the Nile River delta); and (*b*) mixed-energy coasts (coasts influenced by both wave and tidal processes) have short, stubby barriers, numerous tidal inlets, and a backbarrier consisting of marsh and tidal creeks (i.e., Virginia, South Carolina, Georgia, Friesian Islands in the North Sea, Algarve in Portugal, northern New England). Wave dominated coasts are a product of wave-generated sediment transport, whereas mixed energy coasts are a product of both wave- and tide-generated sand transport. The importance of tides in mixed energy coasts is manifested in more tidal inlets, shorter barrier islands, and well-formed ebb-tidal deltas.

Tidal inlets are openings along barrier island chains through which ocean water penetrates to the land thereby providing a connection between the ocean and back-barrier (bays, lagoons, and marsh and tidal creek systems). Tidal currents maintain the main channel of a tidal inlet by continually removing sediment dumped into the main channel by wave action. Some tidal inlets coincide with the mouths of rivers (estuaries), but in these cases, inlet dimensions and sediment transport trends are still governed, to a large extent, by the volume of water exchanged at the inlet mouth (tidal prism) and the reversing tidal currents. Closely associated with tidal inlets are sand shoals and tidal channels located on the landward and seaward sides of the inlets. Flood tidal currents deposit sand landward of the inlet forming a flood-tidal delta and ebb-tidal currents deposit sand on the seaward side forming an ebb-tidal delta (**Figure 13**). The effects of SLR on barriers and tidal inlets are treated jointly in this section because of their complex interactions.

Tidal inlets throughout the world exhibit several consistent relationships that have allowed coastal engineers and marine geologists to formulate predictive morphologic models based on empirical and quantitative studies. Two of these relationships are particularly useful in predicting how tidal inlets will respond to accelerated SLR:

Inlet throat area–tidal prism relationship: The volume of water flowing through a tidal inlet during a half tidal cycle, termed the tidal prism (P, m^3), is closely related to the inlet throat cross-sectional area (A_c, m^2) measured during a spring tide (O'Brien 1931, 1969; Jarrett 1976):

$$A_c = 7.49 \times 10^{-5}\, P^{0.86} \tag{7}$$

Ebb-tidal delta volume–tidal prism relationship: Walton & Adams (1976) showed that the spring tidal prism closely corresponds to the volume of sand (V) contained

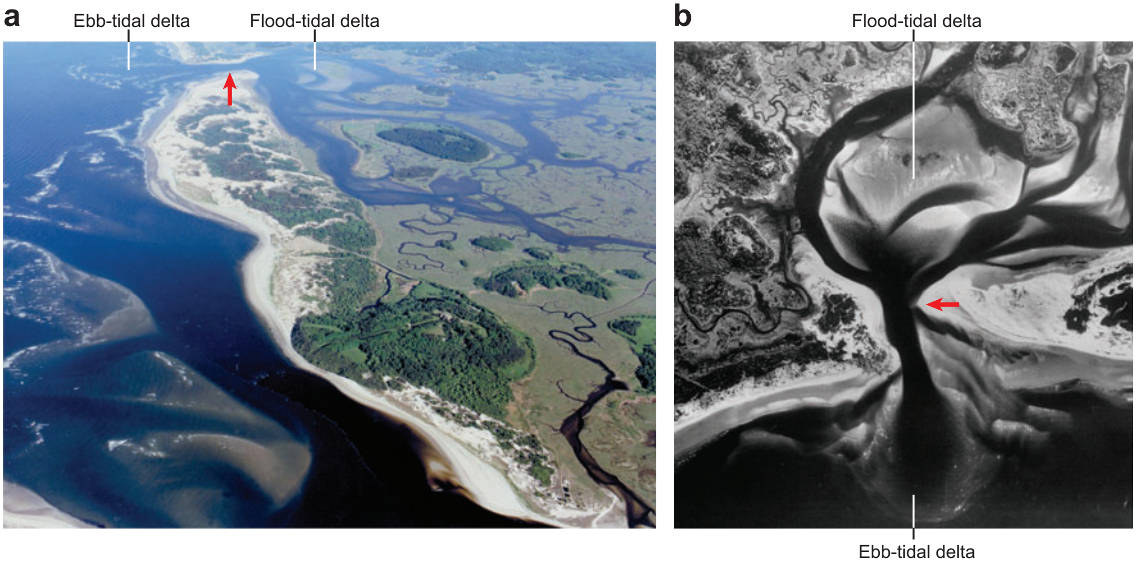


Figure 13

(*a*) Oblique aerial view of the Castle Neck barrier island and tidal inlet system in northern Massachusetts. (*b*) Vertical aerial photograph of Essex Inlet, MA, shown in top of accompanying photograph. Arrows point to the same location in each photo.

in the ebb-tidal delta. Moreover, the relationship improves when wave energy is taken into account:

$$V = 1.89 \times 10^{-5}\, P^{1.23}. \tag{8}$$

Predictive Models of Barrier Response to Sea-Level Rise

The response of sedimentary coasts to sea level changes was discussed in a landmark paper by Curray (1964) in which he related progradation versus retrogradation[3] of a coast to the rate of SLR and whether the rate of sediment supply to the coast results in erosion or deposition. As Curray (1964) showed for tectonically stable margins, the coast retreats in most SLR scenarios unless a high rate of sediment deposition offsets this tendency. Conversely, the coast progrades when sea level falls unless there is a high rate of shoreline erosion. It has been shown that projected shoreline recession or advancement is appropriately determined using a sediment budget approach (Rosati 2005). Given that sediment supply for most barrier coasts is waning or exhausted because of various factors, including the construction of dams on rivers, the building of coastal structures that prevent erosion of bluffs, cliffs, dunes, etc., that would otherwise contribute sediment to the littorial system, and there is little if any sand being transported onshore from the inner continental shelf, an acceleration in SLR

[3] Curray used the terms transgression for the shoreline retreating landward (retrograding) and regression for a shoreline migrating seaward (prograding).

Figure 14

Oblique aerial photograph of Masonboro Island looking south. Long-term erosion of this barrier has led to a continuous thin sand sheet transgressing over the marsh. (Photograph by Bill Cleary).

will exacerbate existing global erosion trends (Pilkey & Cooper 2007). The combined erosion of the ocean and bay side of the barrier leads to an overall narrowing of the barrier.

Long-term effects of storms, negative or reduced sediment supply, and SLR cause beach erosion and eventual destruction of the foredune ridge along barrier coasts. At structured (coastal engineering structures such as sea walls, revetments) or nourished (adding sand to beach from outside sources) beaches, barrier systems may be temporarily held in place, but this can be a costly response and not a permanent solution (Valverde et al. 1999). In an erosional regime, dismantlement of the primary dune is a precursor to barrier denudation, frequent washovers, and the initiation of barrier rollover (Inman & Dolan 1989). This scenario can take place along an entire barrier, such as on Morris Island in central South Carolina and on Masonboro Island along the southern coast of North Carolina (**Figure 14**), or it may involve a segment of a barrier that has experienced preferential erosion and shoreline recession, such as on Nauset Spit along the outer coast of Cape Cod (FitzGerald & Pendleton 2002). Timescales of beach and foredune erosion and initiation of barrier rollover usually encompasses 50 to more than 100 years (Giese 1988) depending on the rate of beach erosion, rate of SLR, frequency of storms, and volume of sand contained in the barrier and its topography (FitzGerald & van Heteren 1999).

In a transgressive regime, the landward transport of sand and onshore migration of barriers occur by three major processes: (*a*) washovers in which sediment is eroded from beach and nearshore/shoreface, transported across the barrier, and deposited along the backside of the barrier forming overwash fans or terraces (**Figure 14**) (McGowan & Scott 1975); (*b*) barrier breaching and tidal inlet formation causing the transport of sand into the backbarrier (Armon & McCann 1979, Leatherman 1979); and (*c*) windblown sand across the barrier, which may occur through dune migration (Jones & Cameron 1977).

The rate of landward migration of a barrier system and the barrier's tendency toward segmentation and possible disintegration depend on the rate of sand loss from the barrier and type of backbarrier setting. For example, the sediment contained in the retrograding Masonboro Island is conserved because the washover sands are being deposited on top of a marsh platform (**Figure 14**). However, once the washovers reach the open water landward of the marsh, the supratidal extent of the barrier will be diminished due to sediment required to fill the water column in the lagoon. Thus, barrier systems migrating across marsh-filled lagoons have a better chance of survival than do barrier system migrating into open water. In regions where the barrier is migrating into a large open sound (e.g., portions of the Outer Banks of North Carolina) the longshore transport rate of sediment along the bayside of the barrier may also be an important factor.

Causes of Barrier Disintegration

The Isle Dernieres represent one of the best-chronicled examples of barrier fragmentation as a function of SLR and diminished sediment supply. This barrier arc is located along the Mississippi River delta coast formed from the reworking of an abandoned distributary lobe of the Lafourche delta complex (Bayou Grand Caillou) (Penland et al. 1988). In the mid-1800s, the barrier arc became separated from the mainland due SLR and high rates of subsidence (Penland et al. 1981). The process of barrier arc detachment from the mainland indicates that the distributary headland no longer supplied sediment to the barrier system (Penland et al. 1988). As depicted in the sequence of maps in **Figure 15**, rapid subsidence in the region (Penland & Ramsey 1990), the lack of new sediment, and sediment losses both alongshore and offshore have combined to thin and segment the barrier. Thus, a robust and continuous barrier chain in the mid-1800s was transformed into a thin, highly discontinuous barrier system by the late 1900s necessitating extensive beach nourishment projects to keep it from disappearing (Kulp et al. 2005). Subsidence is due to dewatering and sediment compaction, but along some coasts (i.e., deltaic coasts) subsidence may also result from tectonic downwarping of the lithosphere.

The high rate of relative SLR for the Isle Dernieres of nearly 1 m century^{-1} (Penland & Ramsey 1990) provides insight into how barrier chains may respond in a regime of accelerated SLR. In other locations, barrier breaching and tidal inlet formation hasten barrier segmentation (Giese 1988). When a barrier is breached, the newly formed inlet becomes a sediment sink as sand is moved seaward and landward, building ebb and flood deltas, respectively (**Figure 13**). This trapping of sand by the inlet depletes the sand reservoir of the littoral transport system, leading to erosion of the downdrift barrier shoreline. The breaching of Nauset Spit along the outer coast of Cape Cod, Massachusetts, during a moderate northeast storm on January 2, 1987, illustrates this condition. In particular, breaching occurred after a long period of beach erosion and barrier thinning caused by a diminished sediment supply and SLR (Friedrichs et al. 1993, FitzGerald & Pendleton 2002). During the breaching process and subsequent widening of the inlet, flood tidal and wave-generated currents moved more than a million cubic meters of sand into the backbarrier, enlarging the flood

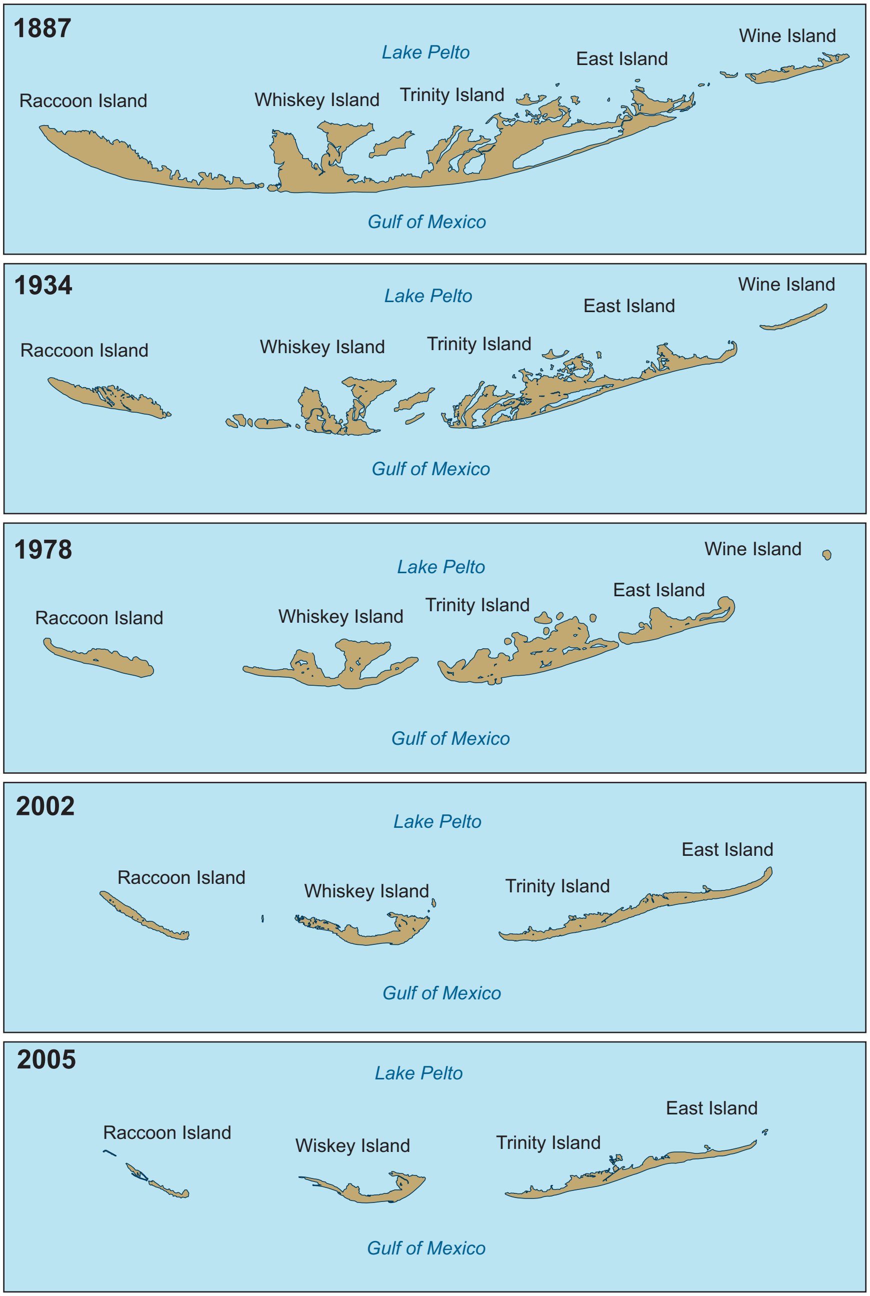


Figure 15

Morphologic evolution of the Isles Dernieres, Louisiana, showing the thinning of the barriers and widening of the tidal inlets (updated maps from Williams et al. 1992).

Figure 16

Sequential photographs illustrating the formation of New Inlet and more recent development of the Nauset Breach, Cape Cod, Massachusetts, which occurred in mid-April 2007. (1985, 1987 photographs from Duncan FitzGerald; 1988 photograph from Joseph Melanson; 2005 photograph from Ted Keon).

deltas and forming extensive intertidal and subtidal shoals (**Figure 16**; FitzGerald & Montello 1993). In addition to sand lost to the backbarrier, the ebb-tidal delta also trapped sand as it grew toward an equilibrium volume (Walton & Adams 1976) by the late 1990s. During this period of sand sequestration at New Inlet, the downdrift shoreline retreated 300 m (FitzGerald & Pendleton 2002).

Tidal Inlet Response to Sea-Level Rise

Relatively little research has been done on the direct effects of SLR on tidal inlets and whether the apparent acceleration in SLR will modify the hydraulics of inlet systems and net sedimentation patterns. Changes in inlet geometry, flow dynamics, and net sand transport directions can be expected if SLR changes the basinal hypsometry (Mota Oliveira 1970, Boon & Byrne 1981, Dronkers 1986, van de Kreeke 1988) and/or the shoaling character of tidal wave propagation into the backbarrier basin (Boon & Byrne 1981, Aubrey & Speer 1985). Perturbation of the tidal wave will alter the filling characteristics of the basin, leading to changes in tidal range, tidal prism, ebb and flood durations, and strength of the tidal currents.

Modeling work of tidal inlets along the West Friesian Islands in the North Sea examined the effects of different SLR scenarios and different sized basins on

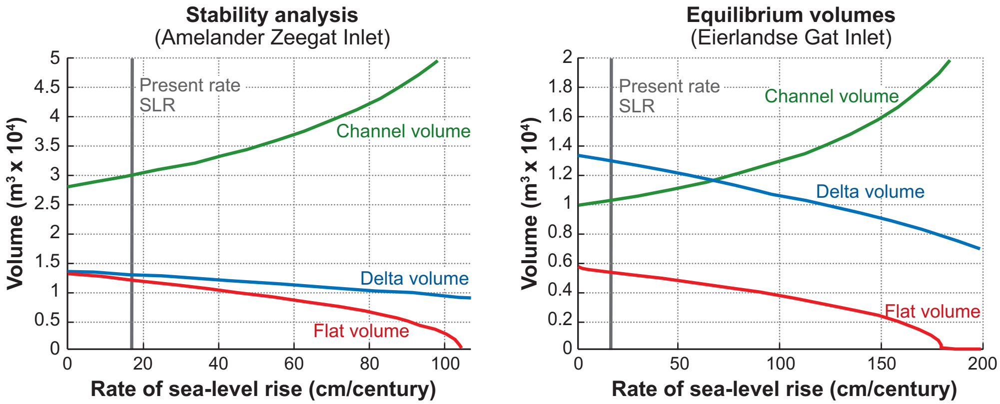


Figure 17

Dynamic equilibrium volumes of tidal inlet elements (see text) as a function of SLR rates for two Dutch inlets. Vertical line represents present conditions (from Van Goor et al. 2003).

accommodation space in the backbarrier basin and the attendant effect on sediment influx (Van Goor et al. 2003, Stive & de Vriend 2004). Using a three-element (ebb-tidal delta, inlet channel, interior tidal flats) sediment equilibrium model and a continuous supply of sand from the adjacent barrier, these investigators showed that for Amelander Gat, inlet channel size increases as the rate of SLR increases, whereas the volume of sand contained in the ebb-tidal delta and tidal flats decrease (**Figure 17**; Van Goor et al. 2003). According to their model, smaller-sized Eierlandse Gat Inlet keeps pace with SLR and maintains its size and extent of its shoals because less sand is required to retain its dynamic equilibrium. Their analyses indicate that relatively high rates of SLR are necessary to overwhelm and drown the inlets. Sediment core data from The Netherlands inner shelf suggest that rates of SLR on the order of 0.8 to 3.0 m per century resulted in the drowning of inlets along the western coast (Beets et al. 1992).

Interactions Among the Backbarrier, Tidal Inlet. Barrier Island

Another approach to forecasting how tidal inlet systems may respond to a regime of accelerated SLR considers the evolution of the backbarrier and how changes to basin hypsometry may affect inlet tidal prism (FitzGerald et al. 2004). Barataria Basin along the central Louisiana coast presents a natural laboratory to study the effects of accelerated SLR on backbarrier systems because it is experiencing one of the highest SLR rates in the United States [0.92 cm year^{-1} for 1947–2006 period (Miner 2007)] and the impacts to the marshlands have been well documented (Barras et al. 1994). The rapid rate of relative SLR is attributed to many natural and anthropogenic causes, which have been addressed by Morton et al. (2005) and González & Törnqvist (2006). The close correspondence in Barataria Bay between relative SLR and wetland loss (**Figure 18**) has been discussed by FitzGerald et al. (2007), who showed that marsh loss was related to several linked processes, including subsidence, marsh front erosion,

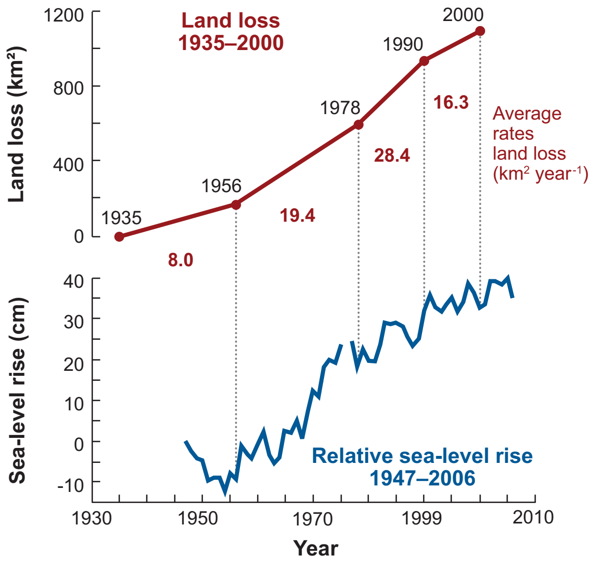


Figure 18

Temporal plots of wetland loss and relative sea level since the mid-1900s (from FitzGerald et al. 2007). Sea-level curve from the Grand Isle gauge (Miner 2007) and wetland loss data from Barras et al. (1994).

and catastrophic scour during large magnitude hurricanes. Long-term conversion of wetlands to intertidal and subtidal environments has steadily increased tidal exchange between Barataria Bay and the Gulf of Mexico, resulting in larger inlet tidal prisms. Direct consequences of the increasing tidal discharge include the enlarging size of tidal inlets (**Table 3**) and growth of ebb-tidal delta shoals (List et al. 1994, 1997; FitzGerald et al. 2004). This growth is exhibited by a 2-km-seaward migration of the 5-m contour at Barataria Pass between 1930 and 1980 (FitzGerald et al. 2004). Since the 1930s, the four tidal inlets that convey tidal flow into and out of Barataria Bay collectively have more than doubled in cross-sectional area in response to 1100 km^2 of marshlands converted to open water (**Table 3**). FitzGerald et al. (2007) have shown that throughout the evolution of this barrier coast, individual inlet cross sections and their combined cross-sectional areas for any one time period fit well within the 95% confidence limits of O'Brien's (1969) relationship, which links tidal prism and inlet cross-sectional area (see Equation 7).

Conceptual Barrier Fragmentation Model

If salt marshes cannot maintain their areal extent through vertical accretion, then these wetlands will be transformed to intertidal and subtidal environments. In mixed-energy

Table 3 Cross-sectional area (m^2) for Barataria tidal inlets (FitzGerald et al. 2007)

Year	Abel	Barataria	Caminada	Quatre Bayou	Total cross-sectional area
1880	0	4304	809	133	5246
1930	395	6271	1353	2590	10609
1980	4193	7182	1532	3777	16684
2006	6669	7374	3372	6726	24141

settings (**Figure 13**), backbarrier salt marshes are characterized by an almost flat topographic profile, generally covering a wide area (width = 2–10 km). This unique hypsometry will lead to a significant increase in the tidal prism, if SLR causes frequent flooding of the supratidal marsh surface or the marshes are eroded. An increase in the tidal range in the backbarrier may also occur as marshlands convert to open water. Increasing water levels will reduce frictional resistance in tidal wave propagation through the inlet, which will result in higher tidal elevations, increased tidal ranges, and enlarging tidal prism (Mota Oliveria 1970). As the O'Brien (1931, 1969) relationship predicts, an increase in tidal prism will strengthen the tidal currents and enlarge the size of the tidal inlet. During the process of increasing tidal exchange between the backbarrier and ocean, the potential for dramatic changes to the inlet shoreline exists. It is well known that tidal prism is the primary factor governing the volume of the ebb-tidal delta (Walton & Adams 1976). Thus, as salt marshes are replaced with open water, enlarged tidal prisms will enable the sequestration of an increasingly greater quantity of sand on ebb-tidal deltas. At the same time, the increase in extent of subtidal and intertidal areas landward of inlets will create accommodation space and promote the formation and/or enlargement of flood-tidal deltas (FitzGerald et al. 2004). As the backbarrier marsh converts to an open-water lagoon, the ensuing changes in inlet hydraulics from ebb to flood dominance will aid shoal development in the backbarrier (Boon & Byrne 1981, Mota Oliveria 1970). To summarize, accelerated SLR may lead to the drowning of marshes, increased tidal prism, and a growth in the volume of sand contained in ebb- and flood-tidal deltas.

In the above scenario, sediment that is deposited on ebb- and flood-tidal deltas and other shoals within the backbarrier comes from sand in the littoral system that enters the tidal inlet from the adjacent barrier islands. This sequestration process may prevent or greatly diminish sand bypassing around the tidal inlet, thereby depleting the sand reservoirs from the adjacent barriers, which may ultimately lead to their fragmentation and onshore migration (FitzGerald et al. 2004). A dramatic example of this process occurred at Ocean City Inlet along the Maryland coast. After the inlet formed in 1935 during a northeast storm, 10 million m^3 of sand was trapped on the ebb-tidal delta during the following 65 years, essentially cutting off the sand supply to the southern downdrift barrier (Stauble 1997). During that 65-year period, the northern end of Assateague Island severely eroded and became highly transgressive, migrating onshore two barrier widths (Williams et al. 2002). Jetty construction and channel dredging following inlet formation exacerbated the erosional condition of northern Assateague Island.

FitzGerald et al. (2004) presented a conceptual model depicting the response of mixed-energy barrier island chains to a regime of accelerated SLR (**Figure 19**). Their evolutionary model is based on marsh conversion to open water causing an increase in tidal prism and growth of the ebb shoals. Sand is also lost from the littoral system as it is moved into the backbarrier to form flood shoals. In particular, flood-tidal deltas will form as the inlets change from ebb- to flood-dominated systems in concert with the transformation of marsh and tidal creeks to open bays. However, this change will not retard the growth of the ebb deltas because that volume depends on the tidal prism.

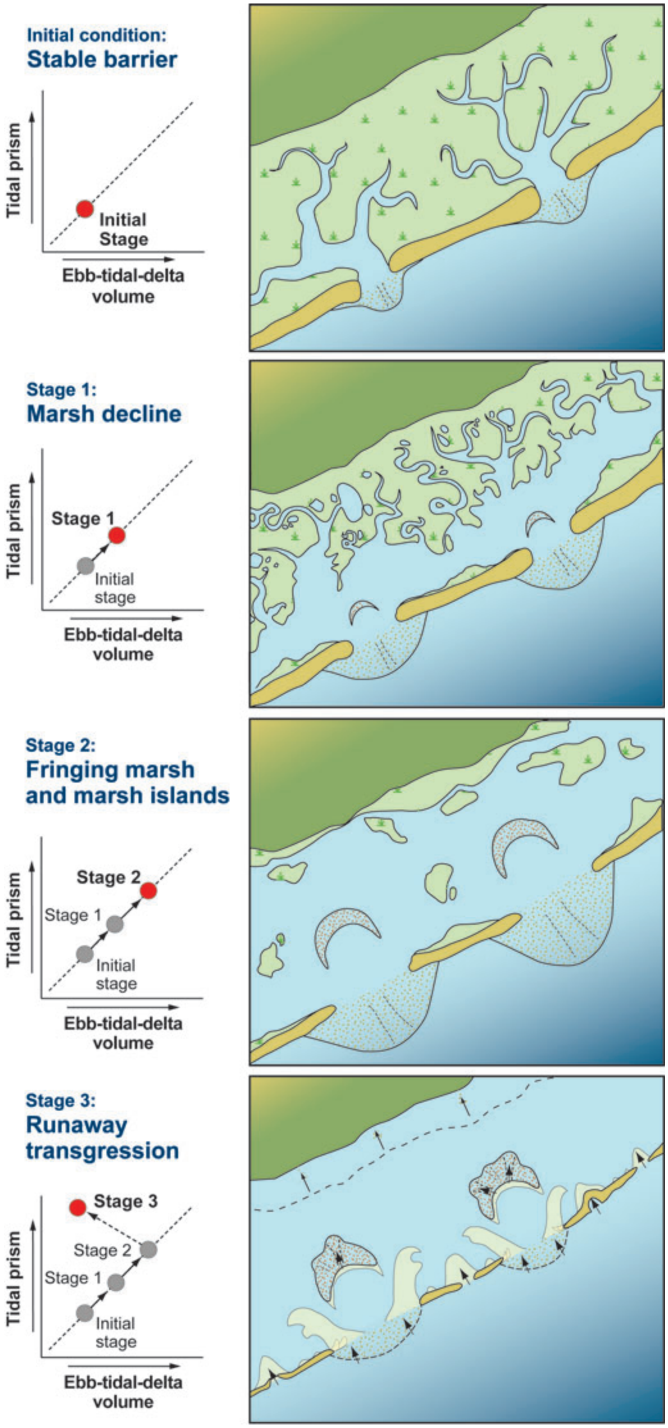


Figure 19

Conceptual evolutionary model of a mixed-energy barrier coast to a regime of accelerated SLR (from FitzGerald et al. 2006).

The capture of sand at inlets and within the backbarrier diminishes sand reservoirs of the barrier islands leading to breaching and eventually retrogradational barriers. The Chandeleur Islands in southeast Louisiana that were largely denuded by Hurricane Katrina (Sallenger et al. 2007) are an example of a barrier arc having reached Stage 3. The rate at which barrier chains may evolve into a transgressive barrier system depends on the future rate of SLR, marsh accretion rates, and the volume of sand contained in the barrier system (FitzGerald et al. 2004).

The evolution of the Friesian Islands illustrates in reverse what can happen to a barrier chain when an alteration of the backbarrier hypsometry induces changes to the tidal prism (FitzGerald et al. 1984). During a 310-year period between 1650 and 1960, the backbarrier area decreased by 30% (149 km^2) due mostly to land reclamation of tidal flat areas along the landward sides of the barriers and along the mainland shore. Secondary losses were attributed to recurved spit extension into the backbarrier. These processes resulted in a reduction of the tidal prism and a coincident narrowing of the tidal inlets by 52% (11 km) and a lengthening of the barriers by 29% (14 km) (**Figure 20**; FitzGerald et al. 1984). The sand supply that was responsible for lengthening the barriers came from a decrease in tidal prism and attendant reduction in size of the ebb-tidal deltas. In a regime of accelerated SLR, if rising waters converted polders (land reclaimed from the sea) back into intertidal flats, the barrier chain would revert to its former morphology due to increasing tidal prism (FitzGerald et al. 2004).

The above examples illustrate that barrier islands, tidal inlets, and backbarrier wetlands are intimately related with respect to sediment reservoirs and erosional and depositional processes. It is also apparent that marshlands of these coasts are particularly vulnerable and may be adversely affected by a regime of accelerated SLR due to their intertidal to supratidal elevation and their limited ability to accrete vertically. This paper has demonstrated that the feedbacks between SLR and dynamic coastal systems will increase the risk to the developed and inhabited United States and

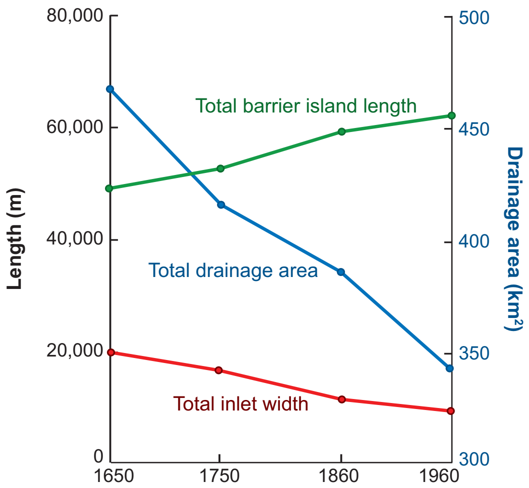


Figure 20

Morphological changes along the Friesian Islands for the 1650–1960 period illustrating how reduction in backbarrier area (drainage area) reduces tidal prism and affects barrier sediment budget (from FitzGerald et al. 1984).

global coasts as well as to habitats utilized by coastal flora and fauna. Consequently, local, state, and federal governments will find accurate predictions of future trends of prime importance when making coastal management decisions.

SUMMARY POINTS

1. The advent of accelerated SLR is an ominous byproduct of global warming, and it effects are likely to be costly both monetarily and in terms of human impact, as more than 10% of the world population lives in coastal regions at elevations lower than 10 m. Many of these regions are susceptible to gradual inundation, storm flooding, saltwater intrusion, and in some regions, devastation by tsunamis.
2. A century of research has shown that, over the long term, SLR causes barrier islands to migrate landward while conserving mass through offshore and onshore sediment transport. Over shorter time periods, additional factors such as sediment supply and the rate of fluid power expenditure control the shoreface profile that translates landward during SLR. Under these conditions, coastal systems adjust to SLR dynamically while maintaining a characteristic geometry that is unique to a particular coast.
3. Future maintenance of salt marshes in a regime of SLR is tied to their ability to import inorganic sediment and produce organic material (particularly root mass). Present estimates of vertical accretion of high-tide marshes indicate that these environments are highly susceptible to drowning if SLR accelerates.
4. In a scenario of accelerated SLR, the transformation of backbarrier marshes to open water will increase tidal prisms resulting in tidal inlet enlargement (leading to erosion of shorelines bordering inlets) and a growth in size of ebb-tidal deltas (extending shoreline erosion for some distance away from inlets). Sand will also enter the backbarrier forming flood shoals. This sequestration of sand on tidal deltas will starve adjacent barrier coasts of sediment, resulting in accelerated erosion.
5. Coastal scientists, engineers, managers, and planners face a great challenge in interpreting the impact of SLR on beaches and barriers because synergistic processes produce responses over a various temporal and spatial scales. Consequently, hydrodynamic and sediment transport modeling is needed for different SLR scenarios to better understand the sustainability of high and low marshes, processes and rates of marsh conversion to open-water areas, and response times for tidal shoal development given different rates of tidal prism increase. Understanding and forecasting the thresholds of barrier and backbarrier stability along both natural and developed coastlines are among the most vital environmental issues and should be the focus of coastal research for decades to come.

DISCLOSURE STATEMENT

The authors are not aware of any biases that might be perceived as affecting the objectivity of this review.

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