REPORT TO CONGRESS: COASTAL BARRIER RESOURCES SYSTEM

Shoreline Change and Wetland Loss in the Coastal Barrier Resources System: A Case Study Analysis

APPENDIX A

U.S. Department of the Interior

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SHORELINE CHANGE AND WETLAND LOSS IN THE COASTAL BARRIER RESOURCES SYSTEM: A CASE STUDY ANALYSIS

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PREFACE

The Coastal Barrier Resources System (CBRS) is the network of coastal barriers protected under the Coastal Barrier Resources Act (CBRA). There are 186 units in the CBRS, ranging in size from less than 20 acres to more than 49,000 acres. Barrier islands, barrier spits and peninsulas, bay barriers, and tombolos are all included in the system. This report presents the results of a 1982 inventory of the CBRS and a study of the shoreline change and wetland loss occurring in selected units of the system.

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INTRODUCTION

The Coastal Barrier Resources System (CBRS) is the network of coastal barriers protected under the Coastal Barrier Resources Act (CBRA). There are 186 units in the CBRS, ranging in size from less than 20 acres to more than 49,000 acres. Barrier islands, barrier spits and peninsulas, bay barriers, and tombolos are all included in the system. This report presents the results of a 1982 inventory of the CBRS and a study of the shoreline change and wetland loss occurring in selected units of the system.

1982 CBRS INVENTORY

The maps on which the boundaries of each CBRS unit are delineated are U.S. Geological Survey (USGS) 7½ minute quadrangle maps dated from the early 1940's through the 1970's. While these maps are sufficient for administrative purposes, they are not a current record of the geomorphology and development status of each unit. In order to develop such a record, an aerial photographic inventory of the CBRS was undertaken in 1982. These photographs were then used to prepare simple habitat maps of the CBRS units showing major natural and humanmade features.

Methods

The U.S. Environmental Protection Agency, Environmental Monitoring Systems Laboratory in Las Vegas, Nevada, gathered the aerial photographs. Aerochrome color-infrared film (Kodak 2443) was used because a light to moderate haze will not obscure the image, the land-water interface on the image is very distinct, and living vegetation can be readily identified because of its false red color. Healthy vegetation displays a red color because the chlorophyll in photosynthesizing plants reflects light in the infrared spectrum. Water appears deep blue on color-infrared images, and sand appears white because all incident light is reflected.

Pictures were taken at an approximate scale of 1:12,000 (1 inch = 1,000 ft) or 1:24,000 (1 inch = 2,000 ft). Most units were photographed in March, April, and May of 1982. A few units (those last delineated by Congress) were photographed in February and March of 1983. All original film was processed, screened, and titled by the EPA Environmental Monitoring Systems Laboratory.

Initial photointerpretation was done at the Environmental Monitoring Systems Laboratory using a Richards single-stage light table fitted with a Bausch and Lomb Zoom 240 stereoscope. The boundary line for each CBRS unit was transcribed from the legal CBRS maps. Habitats and development were interpreted within the boundaries of each CBRS unit and in a 500 ft peripheral zone around each unit.

A simple classification scheme was used in the interpretations (Figure 1). At the time this scheme was developed, the major interest of both the Congress and the Department of the Interior was finding the potentially developable land (i.e., land where homes needing Federal flood insurance might be built). That is why the primary habitat distinction is between wetland and fastland (any nonwetland--defined further below). Buildings and other cultural features were also included to verify the "undeveloped" status of each CBRS unit.

The fastland in each unit includes that portion of the coastal barrier between the mean high tide line on the ocean side and the upper limit of tidal wetland vegetation (or, if such vegetation is not present, the mean high tide line) on the landward side of the coastal barrier. The line drawn for the ocean side of the fastland represents the toe of the dunes (i.e., the line between the vegetated dunes and the backshore of the beach), including dune overwash and blowout areas. In some areas, where pockets of wetland are included in a fastland, a "fastland with interior wetland" classification was used. For these areas, the wetlands were not individually outlined, they were just indicated as a percentage of the total area. For example, an "F/W 10" designation would identify a fastland with about 10% interior wetland. Wetlands include most wetland types except intertidal beaches and bars. Open water was delineated if the water area was larger than 1.4 acres.

Any walled and roofed building other than a gas or liquid storage tank that is principally aboveground and affixed to a permanent site, including a mobile home on a foundation, is defined as a structure. This definition could not be strictly applied in these interpretations because walls are not visible in aerial photographs. All readily apparent structures or groups of structures were delineated. Some roofed but unwalled structures, such as park pavillions, were included in these delineations. Roads, railroads, jetties, docks, and groins were also identified if readily apparent.

A standard land use classification system was used to delineate the peripheral zone around each CBRS unit. Developed areas, undeveloped areas, agricultural fields, wetlands, open water, and roads were identified.

Interpreted photographs were sent to Martel Laboratories, Inc., in St. Petersburg, Florida, for map-making. Martel checked the photointerpretation and CBRS boundaries and added an interior open water classification. This class includes all water bodies that are completely enclosed by fastland or wetland, such as ponds. Maps were prepared as mylar overlays to 1:24,000 USGS topographic quadrangles except in Massachusetts where the standard large scale USGS map size is 1:25,000.

The U.S. Fish and Wildlife Service (FWS) received these maps and checked them for quality and accuracy. The maps were then digitized using FWS' computer geographic information system and following standard operating procedures (U.S. Fish and Wildlife Service 1982). Acreage statistics for each CBRS unit were computed using this digitized data.

Symbol	Legend	Symbol	Legend
			COASTAL BARRIER INTERPRETATION
FL	Fastland		Structure and associated developed area
WL	Landward wetlands including tidal flats (between fastland and open waters)	$\boxed{13}$	Concentrated structures and associated developed areas (number represents total count)
F/W5	Fastland with interior wetland (number represents approximate percent of interior wetland--in this example, 5%)		Jetties, docks, groins
IOW	Interior open water, water totally enclosed within the barrier fastland or wetland		Road
			Study area boundary
OW	Open water		
			PERIPHERAL LAND-USE INTERPRETATION
1	Developed (includes residential, industrial, recreational)	5	Open water
2	Undeveloped (includes open space)		Road
3	Agriculture		Limit of interpretation
4	Wetland		

Figure 1. Interpretation classes used in the Coastal Barrier Resources System inventory.

Results and Discussion

Acreage statistics for each CBRS unit were compiled into State tables (see Tables 1 to 15, pages 118-129). A summary of the total acreages by State is presented in Table 16 (page 130). The digital maps of the CBRS, housed with FWS in Slidell, Louisiana, and the descriptive statistics in these tables are intended to serve as baseline data characterizing the CBRS.

Probably the most useful statistic in the tables is the total acreage without water (total w/o water). This number represents the total land area, including wetlands, in the unit. If land area is lost through erosion or storm damage, this number will decrease. The number of structures is also important. It is an index of the level of development present in the unit when the CBRS was created. This number can be compared to later figures to see how much development has occurred in the unit.

Some care should be taken when quoting habitat areas for selected units. Mudflat and sandflat boundaries (wetland in our simple habitat classification system) were difficult to delineate in some areas because small changes in water level covered or uncovered substantial acreages of these broad flats. In some units in south Texas where large expanses of intertidal flats occur, what is classified as wetland on one day may appear as open water the next.

SHORELINE CHANGE AND WETLAND LOSS ANALYSIS: PREPARATION FOR THE 1987 REINVENTORY

The CBRA mandates that the CBRS be reinventoried at least every 5 years in order to update the official CBRS maps. Section 4(c)(3) states:

> The Secretary shall conduct, at least once every five years, a review of the [CBRS] maps...and make, in consultation with appropriate officers...such minor modifications to the boundaries of system units as are necessary solely to reflect changes that have occurred in the size or location of any system units as a result of natural forces.

The first such review should occur in 1987.

As background for conducting a meaningful review of the CBRS maps, FWS undertook a study of the historic geomorphological change in selected units of the CBRS. The primary objective of this study was to understand the dynamics of the barriers and to determine where land loss problems might be greatest in the CBRS. A secondary objective of the study was to determine whether this land loss is being caused by natural processes, human manipulations of the coastal environment, or both.

STUDY APPROACHES

Two related approaches were used in this study. First, case studies of 18 coastal barriers containing CBRS units were conducted. The case study areas were selected to cover a wide geographic area, a diversity of geologic types of barriers, and a range of development pressures. Because these case studies were performed using information available in the scientific literature, it was also necessary to make selections based upon a past history of research in the area.

Each case study describes the geologic history of the area surrounding the CBRS unit and the coastal processes that are important in shaping the barrier. Any shoreline or wetland change that is occurring is discussed and the causes for the changes are identified. The development status of the coastal barrier and its value to fish and wildlife are also briefly discussed where that information is available. Finally, recommendations regarding the long-term stability of the barrier and the conservation of the unit are presented.

In-depth studies of entire coastal barriers are the only way to understand what is happening or is likely to happen to a particular CBRS unit. While some CBRS units encompass whole barriers, most CBRS units are just small pieces of much larger coastal barrier systems and the natural processes that control their existence operate at that larger scale. Focusing narrowly on just what is happening within each unit's boundaries might result in overlooking impending alteration or misunderstanding the causes of change. The simplest illustration of this is an area where a jetty or groin is built outside, but upcurrent, of a coastal barrier unit. The sediment starvation such a jetty would cause downdrift would result in rapid erosion of the unit shoreline. The human cause of this erosion (the jetty) would not be recognized unless the larger area were considered.

The case study analysis covers 10% of the CBRS. It is very difficult to draw conclusions about the entire CBRS based on such a small sample size. Therefore a second approach was taken to obtain a smaller amount of data about a larger number of CBRS units. Twenty-seven CBRS units were selected for this analysis. The 1982 digital maps of these units were compared to older maps to look for changes in the land mass of the units.

Older maps were obtained from two sources. The FWS database at the National Wetlands Research Center (NWRC) in Slidell, Louisiana, already contains a large number of digitized habitat maps from the 1950's, including much of the gulf coast and selected areas of the Atlantic coast. Any of these maps that covered CBRS units were used. The rest of the historic maps were obtained from USGS. They were older edition topographic maps prepared in the 1930's, 40's, or 50's. These maps were digitized at FWS as described previously.

Historic maps were compared to 1982 CBRS unit maps using NWRC's computer geographic information system (GIS). The boundaries of each CBRS unit digitized into the 1982 maps were overlain onto the historic maps by computer to define exactly the same area in both maps. The acreage of land existing in 1982 was then compared to the acreage at the previous date to look for land loss. These numerical comparisons do not provide information about how and why change is occurring, but they do show where and to what extent it is occurring.

CASE HISTORIES OF EIGHTEEN COASTAL BARRIERS

Case studies of the 18 coastal barriers (19 CBRS units) that were selected follow, starting with Plymouth Bay, Massachusetts, and working down the Atlantic coast and around the gulf coast to Boca Chica, Texas (Figure 2). These case studies contain regional and site-specific information about the physical characteristics of the 18 coastal barriers. We do not discuss general characteristics of all barriers. General information about types of coastal barriers, waves, tides, sea-level rise, storm impacts, sediment sources, and the like is presented in Chapter 2, Volume 1 of this CBRS report. Those readers who are not familiar with these topics may wish to read that chapter before these case studies.

CBRS UNIT C04--PLYMOUTH BAY, MASSACHUSETTS

Geomorphology

The first major indentation along the Massachusetts shoreline south of Boston is Plymouth Bay. Located 31 mi south of Boston and 19 mi west of Cape Cod, Plymouth Bay is rimmed by barrier spits, baymouth bars, embayment muds, and sand, gravel, or cobble beaches that coalesce with glacial deposits of Wisconsinan age (late Pleistocene).

Plymouth (Long) Beach and Saquish Neck are barrier spits flanking the bay. Duxbury Beach is a tombolo formed by the growth (progradation) of spits from Gurnet Point and Brant Rock (Figure 3). These features have evolved from the extensive reworking of glacial deposits by marine processes during the Holocene. The sediments at the ends of the spits have been reworked into shoals which are bisected by tidal channels. These channels extend landward into Plymouth Harbor through a well-developed mudflat-tidal marsh deposit which overlies glacial lake clays and silts. Sediments on the spits commonly range from fine to medium sands with grain size increasing towards the major sediment source areas. Concentrations of gravel are found close to Gurnet Point. Sediments updrift of Plymouth Beach (towards Rocky Point) commonly range from cobble to boulder size. The origin of these deposits is the nearby bluffs composed of glacial till. These glacial deposits serve as the principal source of beach sediments for the entire area.

The Plymouth Bay CBRS unit comprises the northern third of Plymouth Beach and the eastern third of Saquish Neck (from the Coast Guard Station at Gurnet Point westward to where cultural development begins).

Geologic History

The landscape of this area is heavily dominated by the effects of Wisconsinan

Figure 2. Location of the 19 case study CBRS units.

glaciation. The Plymouth Bay area is surrounded by moraine (accumulated earth and stone) and other glacial deposits. The outwash from these deposits contributed to the formation of glacial lake Cape Cod. South of Manomet, a moraine borders the coastal zone. This feature largely consists of boulder-sized material which grades into sand and gravel outwash deposits in the vicinity of Plymouth Harbor. Plymouth Bay is underlain by silts deposited in the glacial lake Cape Cod basin. To the north, Gurnet Point and Saquish Neck are drumlins (loosely consolidated glacial deposits) that exist amidst more resistant glacial lake clays and glacial stream sands and gravels. Differential erosion of the glacial lake clays and Holocene sea-level rise are the major factors accounting for the sizable indentation of the coastline at present day Plymouth Bay.

Modern Changes

A history of shoreline changes for the area is limited but it does suggest that the features in southern Plymouth Bay evolved through the erosion of the bluffs at Rocky Point (Figure 3). Undoubtedly, some additional sediment bypasses Rocky Point and contributes to the overall system, but no information exists for a precise definition of the sand-sharing system.

Shoreline change records from the U.S. Army Corps of Engineers (1959) and the Massachusetts Coastal Zone Management Office (1985) indicate that net retreat occurred between 1853 and 1951 at all but Gurnet Point and the tip of Saquish Point (Figure 4). Maximum net erosion over this almost hundred year period at Duxbury Beach and the southwest-facing side of Gurnet Point is about 50 ft. The south-facing shoreline of Saquish Neck underwent marginal change, whereas the north shore eroded up to 280 ft. The CBRS unit area of Saquish Neck experienced a similar trend.

Plymouth Beach spit is currently undergoing transgression although it has been largely progradational over the past 130 years. Data from 1853 through 1951 indicate that the shoreline had migrated seaward at all but the terminus of the spit (Figure 5). In a subsequent 23-year period from 1951 to 1973, however, a net landward displacement of 65 ft occurred at the updrift end extending to the midportion. This reversal may be related to the initial construction (1940) of the updrift Rocky Point revetment. South of the spit, the shoreline retreated 230 ft with a concomitant steepening of the offshore over a 100-year period (1860-1959). During the same 100-year interval, the shoreline at Rocky Point eroded 656 ft (U.S. Army Corps of Engineers 1962).

Net changes in the shoreline position during the last 130 years include net erosion of the north- and east-facing shorelines along Duxbury Beach and Saquish Neck, net deposition at the terminus of Gurnet Point and Saquish Neck, very little net change along the south-facing shoreline west of Gurnet point, transgression but little change in width of the terminus of Plymouth spit, and recent retreat of the oceanside and bayside shorelines extending from the headland to the midpoint of the spit. That portion of Plymouth Beach that is in CBRS has been accreting since 1950. This coincides with the erosion of the updrift end of the spit during the same time period and the overall net erosion of the entire spit (see Management Implications section).

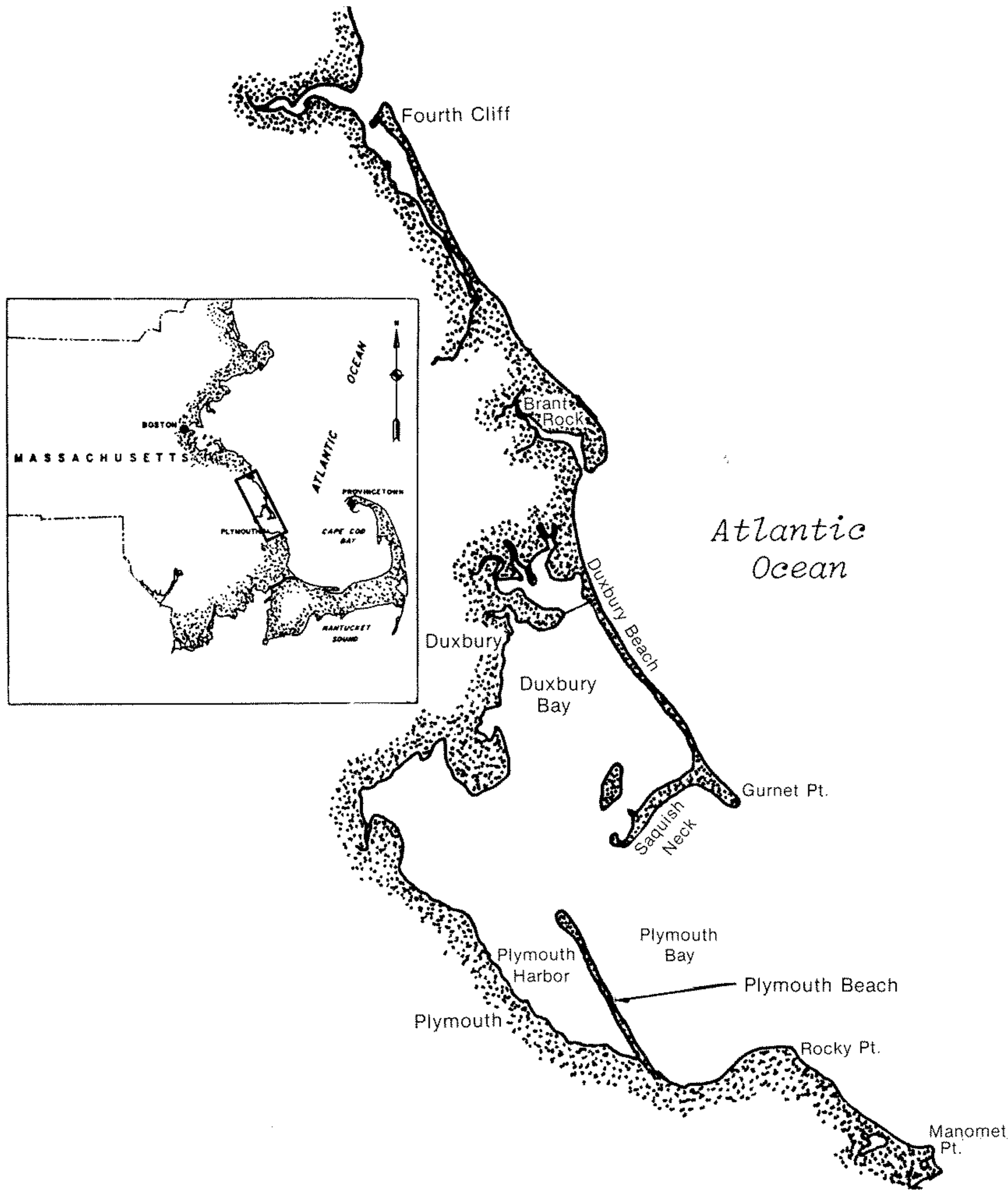

Figure 3. Location of Plymouth Bay, Massachusetts (from U.S. Army Corps of Engineers 1962).

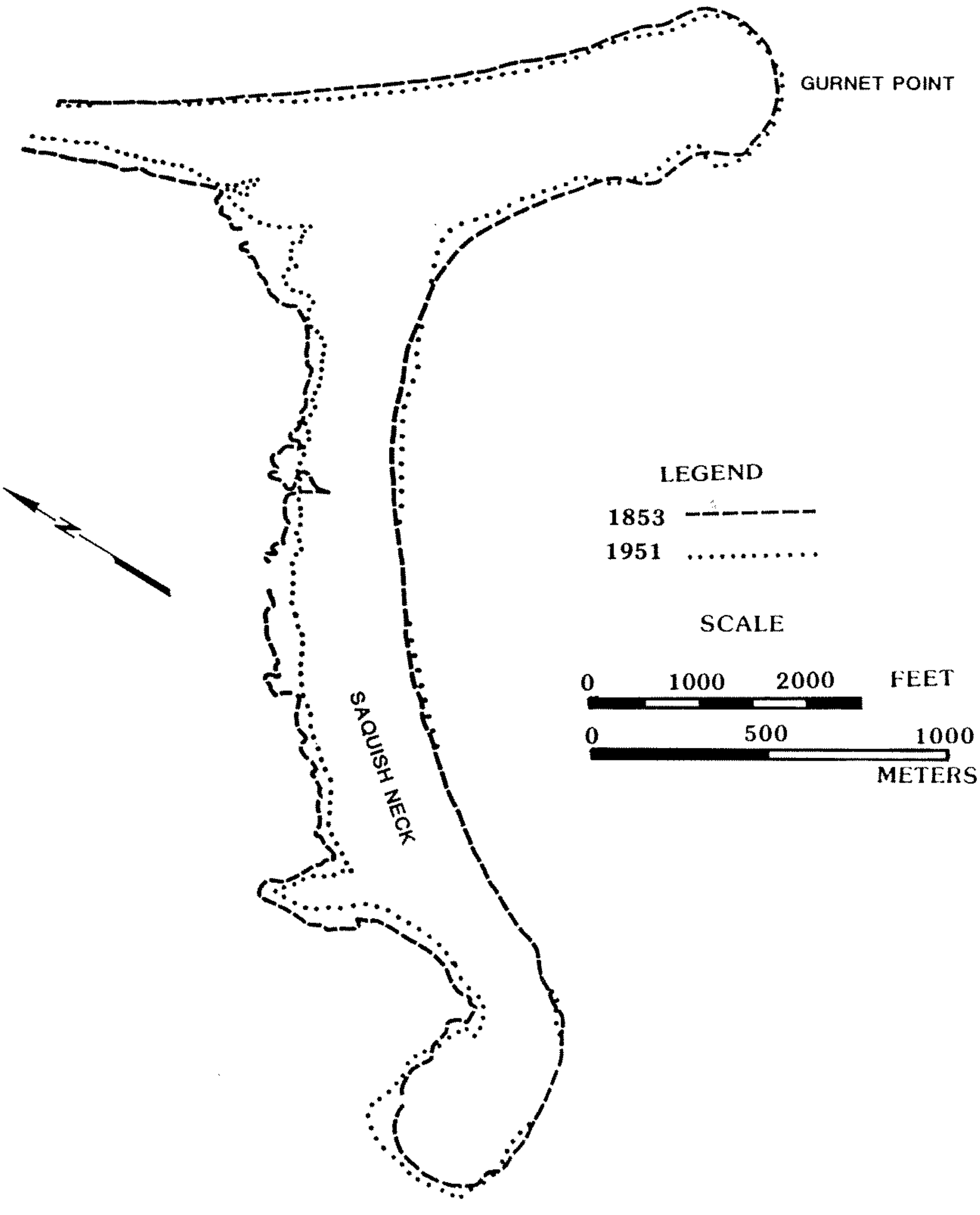

Figure 4. Recent shoreline changes, Duxbury Beach, Massachusetts. Discontinuities in lines represent areas where the map or photo bases were inadequate for accurate delineation (Massachusetts Coastal Zone Management Office 1985).

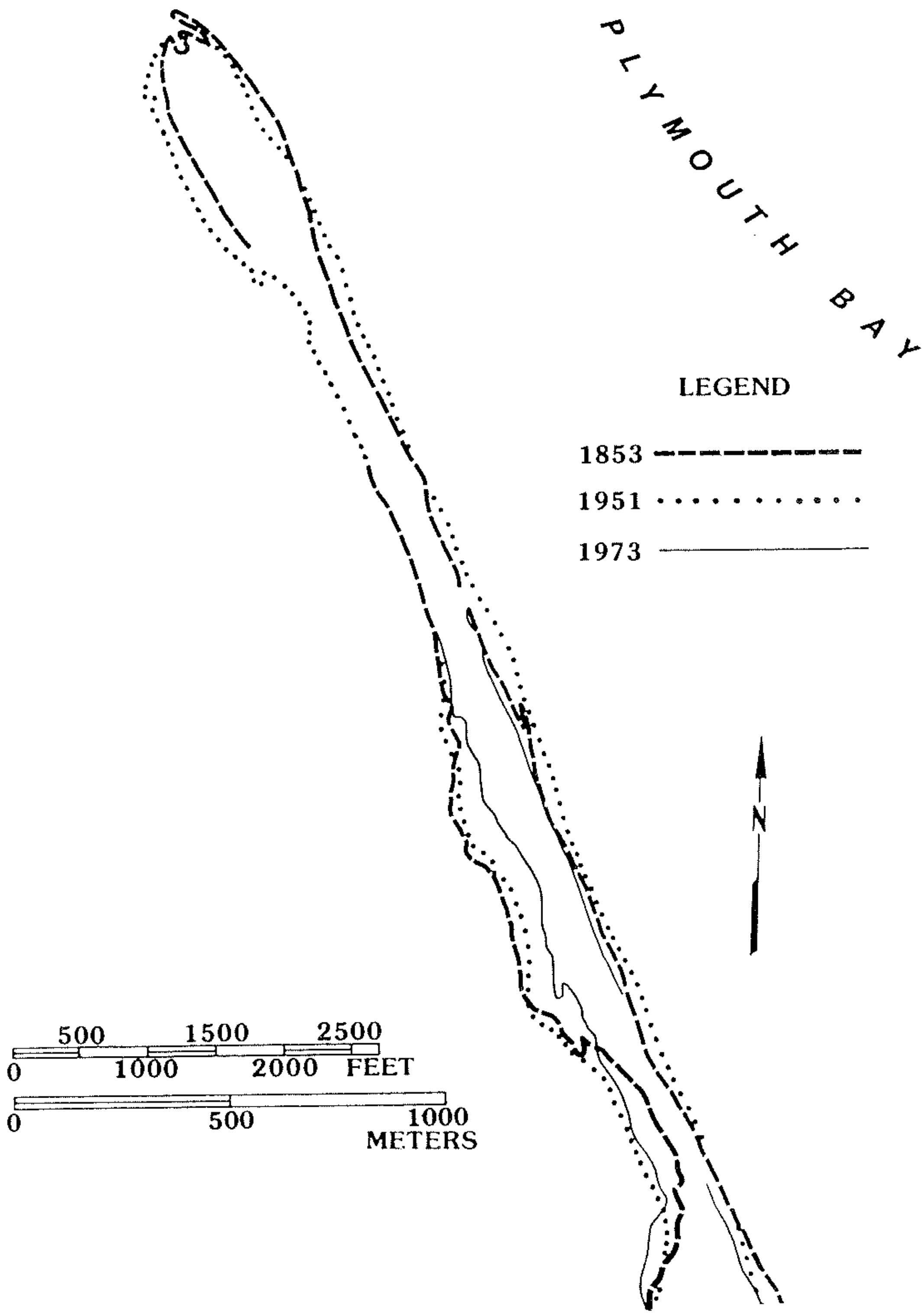

Figure 5. Recent shoreline changes, Plymouth Beach, Massachusetts. Discontinuities in lines represent areas where the map or photo bases were inadequate for accurate delineation (Massachusetts Coastal Zone Management Office 1985).

With initial settlement of Plymouth Harbor and subsequent development of residential, commercial, and recreational facilities in the area, efforts were made to limit erosion in Plymouth Bay. Residences located along the bluff at Rocky Point (updrift of the Plymouth Beach part of the CBRS unit) were fronted by a 0.9 mi long revetment designed to curtail erosion (constructed as a town project in 1940 and reconstructed and lengthened by the State in 1960-68). Farther downdrift at the base of the spit, groins (1968) and a seawall (1961) were constructed in front of a bath house facility (still updrift of the CBRS unit). The principal structures on Plymouth Beach include an experimental adjustable groin (1970) located midway along the oceanside shoreline, a 164 ft long jetty designed to limit shoaling in the Plymouth Harbor channel, and a rubble dike located centrally along the main axis of the spit (construction dates were orally provided by the U.S. Army Corps of Engineers, Waltham, Massachusetts, and the Plymouth Town Engineer).

Coastal Change Processes

Plymouth Bay is subject to tides and locally generated wind-waves. This is a mesotidal (7-13 ft tidal range) environment. The semi-diurnal tides have a mean range of 9.8 ft and a spring range of 11.8 ft. Wave and storm surge data for this area are unavailable; however, records from Boston Harbor indicate that storm-generated water levels can exceed 9.8 ft above the normal high tide line (U.S. Geological Survey 1978). Fetch length is greatest from the northeast and lessens to the south and east because of the presence of Cape Cod. Greatest impacts on the shoreline are from hurricanes and extratropical storms (northeasters). The frequency of northeasters exceeds that of hurricanes. The most vulnerable areas are Duxbury Beach and the Manomet Point headland. Plymouth Beach is less susceptible to storm impacts because of the sheltering effect of Duxbury Beach and Gurnet Point to the northeast and the dissipative effects of a low angle shoreface.

Plymouth Beach spit is a relatively low-lying feature that attains a maximum elevation of 20 ft above the mean sea level but is generally less than 10 ft above mean sea level. The littoral drift rate has not been established but it appears to be very low. Shepard and Wanless (1971) compared historical navigation charts and aerial photographs over a 190-year period and concluded that very little change occurred in the configuration of the tidal channels and Plymouth Beach. Further, the shoreline from Rocky Point to the end of Plymouth Beach has been stabilized somewhat with a seawall, revetment, bulkhead, and groins, thus reducing the otherwise low drift-rate.

Apparent (based on tide-gauge data) sea-level rise is an important long-term component of erosional processes. Several tide gauges in the region indicate a sea-level rise of approximately 1.2 inches/decade (3 mm/yr) (Hicks 1978).

Management Implications

This CBRS unit is composed of two subunits--Saquish Neck and Plymouth Beach. Although both are located within the Plymouth Bay system, most of the cultural forces that affect one subunit have no impact on the other. Plymouth Beach and Saquish Neck therefore should be managed as separate units.

The U.S. Army Corps of Engineers (1959, 1962) has paid considerable attention to southern Plymouth Bay in response to difficulties with navigation and shoreline erosion. The modifications previously mentioned have resulted in stabilization of the shoreline in some places, but the long-term geomorphic responses and resulting management implications have not been considered. The natural system is undergoing net erosion in the principal sediment-source areas. The spits and tombolo have been eroding rapidly, particularly along the midportions of Plymouth Beach and Duxbury Beach. That part of Plymouth Beach that is within the CBRS has actually been accreting. This is contrary to the trend for the entire spit but is not unusual in that the CBRS section of the spit is at the terminal (downdrift) end.

A management strategy should predict the magnitude of shoreline change both within and downdrift of the stabilization structures. For example, the goal of protecting residences from erosion on the bluff and along southern Plymouth Beach has been realized through the construction of a revetment, seawall, and groin system. However, maintaining the integrity of the updrift end of the spit (the CBRS unit) to the lee of the groins and providing a viable beach for recreational purposes may pose a problem in the longer run. It is difficult to predict the long-term effects of the shore protection structures because there is no qualitative information on littoral drift rates and the sand-sharing system has not been completely delineated. Identifying localities where sands of suitable size can be retrieved to nourish critically eroding shorelines would be useful.

The Plymouth Bay area has tremendous historical significance and supports a broad range of land uses along its shorefront. The development pressure for both Plymouth Beach and Saquish Neck is high. Many existing structures (outside of CBRS but within the same sediment-sharing system) are close to the shoreline and, thus, erosion is a critical problem. The U.S. Army Corps of Engineers has already expended considerable effort and monies for erosion protection. Continued development will lead to further pressures for Federal assistance for erosion protection.

The fish and wildlife value of the area lies principally in the broad tidal flats surrounding both Plymouth Beach and Saquish Neck. There is also some salt marsh on the lee side of Saquish Neck within the CBRS. Pollution is the principal negative impact of cultural development on these tidal flats and the marsh.

CBRS UNIT C17--POPPONESSET SPIT, MASSACHUSETTS

Geomorphology

Located centrally along the south shore of Cape Cod, Popponesset Spit progrades to the northeast into Nantucket Sound. This is a recently formed feature with a northeast-southwest orientation. A marine bluff approximately 40 ft high forms the shoreline to the southwest. The bluff is composed of Pleistocene outwash deposits which provide material for the beaches to the north. Net direction of littoral drift along this part of the coast is clearly to the northeast, and the existing configuration of the spit can be attributed to persistent episodes of spit accretion and breaching within an overall framework of landward retreat (Figure 6). The CBRS unit includes the entire spit and extends bayward, incorporating Little Thatch Island and surrounding mud flats. The entire unit, however, only comprises some 82 acres, of which only 15.9 acres are emergent land.

Geologic History

The effects of Pleistocene glaciation have been widely chronicled for Cape Cod, which is believed to have formed during the late Wisconsinan (Chamberlain 1964; Strahler 1966). Popponesset Spit occupies the central coastal portion of a massive outwash plain that forms the southwestern portion of Cape Cod. Source material for this deposit is the moraines which converge 12.4 mi north of Popponesset Spit. The glacial deposits in the Succonnesset headland to the southwest provide sediment for the barrier spit. The beaches are composed of medium- to coarse-grained sand derived from these deposits.

Modern Changes

Numerous changes in the length and seaward extent of Popponesset Spit have occurred during the last two centuries. An analysis of historical maps and charts for the last two centuries and an interpretation of aerial photographs for a 30-yr period were conducted by Aubrey and Gaines (1982). They indicate that the spit has varied in length from a maximum of 1.7 mi to its present length of 0.6 mi during a 30-yr timespan (Figure 7). Variations in spit length can be attributed to breaching which is a chronic condition at Popponesset Spit. Breaches have occurred at four locations. For the most part these are at the proximal end of the spit, a common occurrence in barrier spits. The only exception was a breach formed in 1954 in response to Hurricanes Carol, Edna, and Hazel at the midpoint of the spit when its length was also at its greatest (Figure 7, November 1955 panel). The breach still exists but now represents the approximate terminal end of the spit. Since 1961 the terminus has widened and lengthened, and the axis of the spit has shifted counter clockwise. Despite this accretion at the terminus, the entire spit has experienced transgression with annual rates varying from 4.9 to 11.5 ft. The average width, however, was found to be constant, indicating a net westward movement of the spit. Over a longer period of time (1890-1978) the spit has shifted westward (landward) some 475 ft (Figure 6; Aubrey and Gaines 1982).

Coastal Change Processes

Popponesset Spit occupies a portion of Nantucket Sound with a mean tidal range of 2.5 ft and a spring range of 3 ft (U.S. Dept. of Commerce 1985). The strongest winds originate from the east, whereas prevailing winds are westerly.

The effect of wind-generated surges can be severe. The maximum recorded storm surge elevation for this region was 10 ft above mean sea level (Weigel 1964). Storm activity results in rapid erosion of shore bluffs, increased potential for breaching, and increased frequency of overwash. The previous discussion

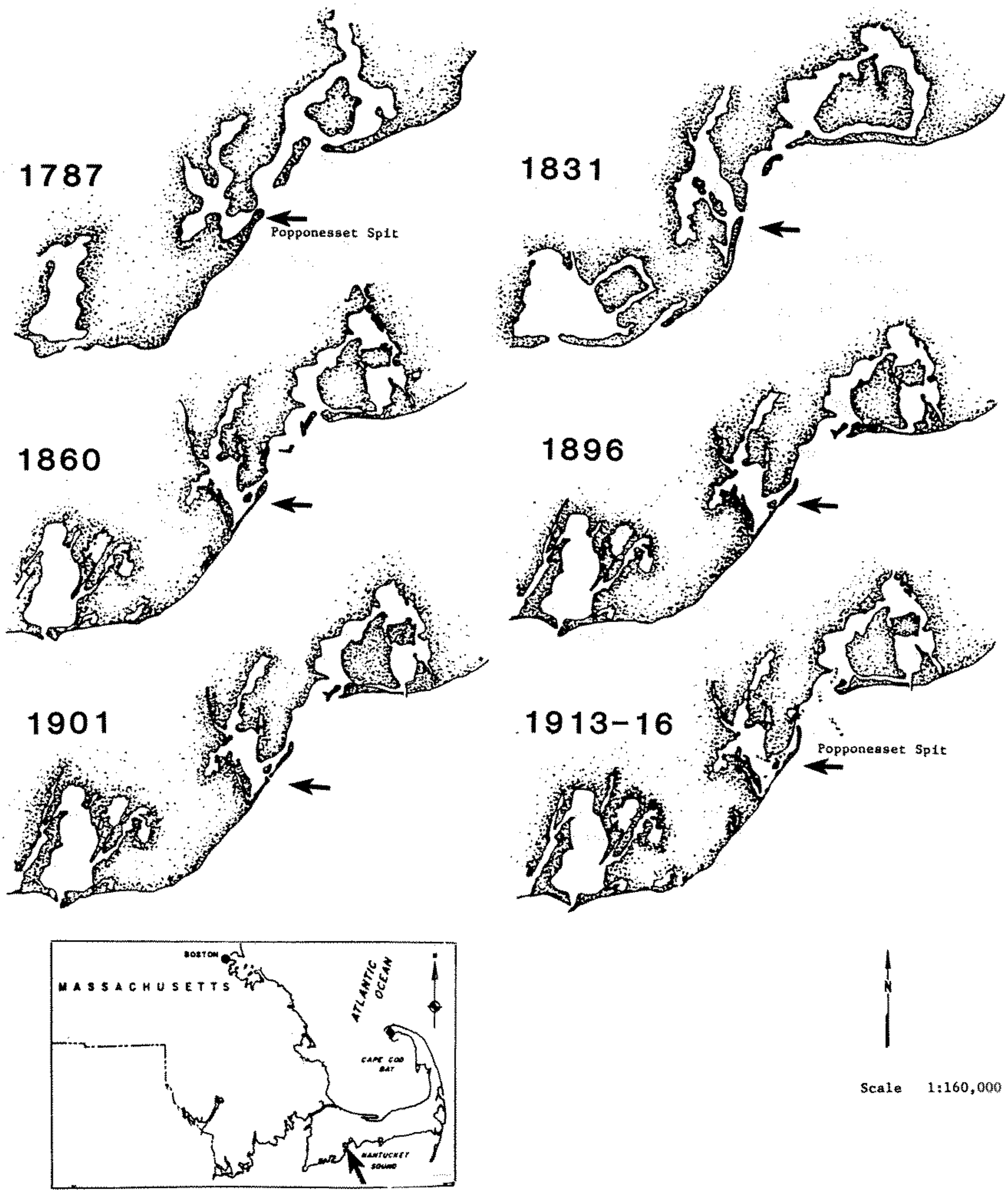

Figure 6. Changes in configuration and position of Popponesset Spit, 1787-1916 (Aubrey and Gaines 1982).

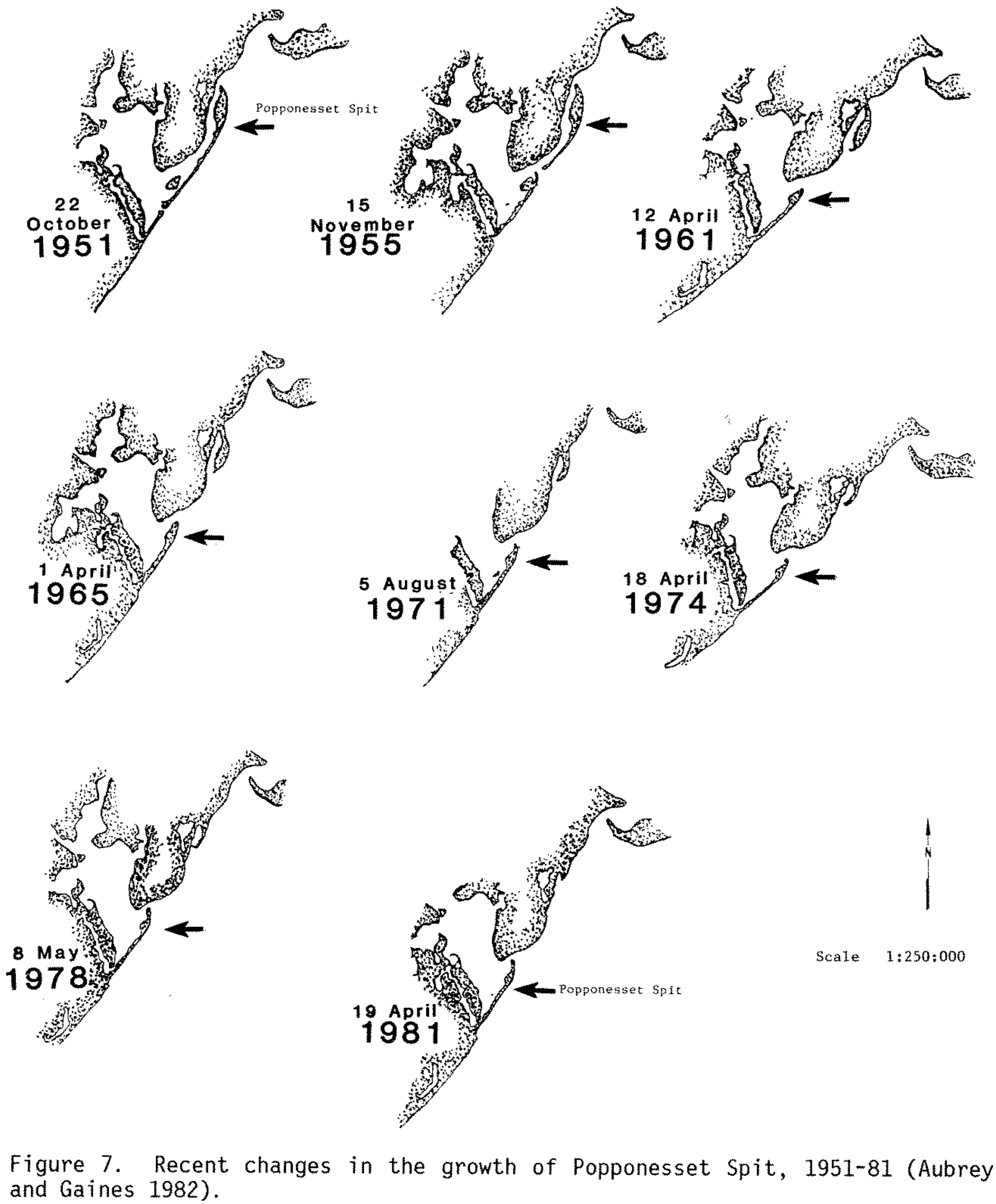

Figure 7. Recent changes in the growth of Popponesset Spit, 1951-81 (Aubrey and Gaines 1982).

of breaching by Hurricanes Carol, Edna, and Hazel indicates the conspicuous effect of hurricanes.

Net longshore transport for the spit is to the northeast; however, the transport direction does undergo reversals. Aubrey and Gaines (1982), in a 1-yr study, indicated that northward transport occurred in the spring with a reversal in the fall. The remaining months exhibited variable drift directions. The total littoral drift rate for Popponesset Spit is small, owing to the presence of relict inlet channels and the lack of sediment bypassing. The relict inlet channels appear to be littoral sediment sinks.

Apparent sea-level rise, which is an important long-term aspect of erosional processes, is approximately 1.2 inches per decade (3 mm/yr) during this century (Hicks 1978).

Management Implications

Attempts to stabilize the Popponesset Beach shoreline began during the early 1950's with the placement of rock groins beyond the proximal end of the spit (outside and updrift of CBRS). A localized effect of these structures was the downdrift starvation of the spit and increased erosion, including the near formation of an inlet at the proximal end of the spit. Sand mining and dredge spoil disposal are other activities in the area. Although the records are incomplete, estimates suggest that 850,000 yd^3 of sediment were dredged from the Popponesset area from 1916 to 1982 for State and private navigation projects (Aubrey and Gaines 1982), and dredging is still occurring.

Any management strategy should work towards the effective placement of dredge spoil. Areas where narrow and low beaches exist or where beaches are updrift of the net littoral transport direction are primary locations for beach nourishment projects. Because the shoreline is persistently migrating landward in response to sea-level rise and major storm events, careful consideration of design criteria for groins and other structures is required; groins built too high and long will serve no useful purpose. The deployment of a seawall without periodic beach nourishment will not provide a recreation opportunity in the long run. Many shorefront property owners updrift of the CBRS unit have constructed seawalls and revetments. Barrier spits such as Popponesset are very vulnerable to breaching, especially at the updrift end. The potential for an inlet to form in the vicinity of Popponesset Island is high. Continued construction of shoreline protection structures updrift of the CBRS without beach nourishment will increase the probability of inlet formation within the CBRS unit.

CBRS UNIT F10--NAPEAGUE BEACH, NEW YORK

Geomorphology

The south shore of Long Island, New York, can be generally classified as a headland-barrier island complex. Its eastern end primarily consists of marine headlands or bluffs formed from erosion of late Pleistocene glacial deposits. Napeague Beach (Figure 8) is an intervening zone of deposition that is believed to have evolved from the merging of spits that originated from two headlands (one west of Napeague Bay and one to the east at Hither Hills) (Taney 1961; McCormick 1973). The headlands have considerable relief which often exceeds 65 ft and the shoreline exhibits numerous irregularities. Beaches fronting the headlands are narrow and are composed of cobble to boulder size materials which eroded from the headlands. In comparison, Napeague Beach consists of a low sandy beach that is backed by a series of linear dunes that range from 10 to 20 ft in height. The CBRS unit only occupies a small section of Napeague Beach, from the central portion of Napeague Harbor southward to the Atlantic shoreline and extending east to west for a maximum distance of 1 mi. A salt marsh occupies the back one-third of the CBRS unit and fringes the western margin of Napeague Harbor.

Geologic History

The eastern end of Long Island is dominated by the effects of Pleistocene glaciation. The headland to the east of the CBRS unit is a glacial moraine; the headland to the west consists of both a moraine and an outwash plain. During the Holocene marine transgression these areas were reworked to form both Napeague Beach and the barrier island system to the west. Interbedded marine sands, gravel, and clay underlie the Pleistocene deposits.

Modern Changes

A considerable body of information exists concerning shoreline change and net littoral drift characteristics for the entire south shore of Long Island. Southside shoreline changes between 1834 and 1979 indicate that net recession has occurred between Montauk Point and Napeague Beach (Figure 9). A lack of data for the shoreline west of Napeague Beach prevents any interpretation of net shoreline response for the entire headland section. However, at Napeague Beach, the initial interval between 1834 and 1892 exhibited the greatest amount of erosion, which ranged from 2 to 11 ft/yr. This was followed by net accretion of up to 4.2 ft/yr along two-thirds of the entire headland shoreline during the next period between 1892 and 1933. During the final interval between 1933 and 1979, sections of Napeague Beach accreted up to 1 ft/yr but the beaches flanking it receded from 1 to 5 ft/yr, giving rise to another interval of net erosion along the headland section. Despite recent episodes of accretion, the resulting net long-term change points to erosion of Napeague Beach. Further, a retreat rate of 1 ft/yr has been estimated for the headlands (McCormick and Toscano 1981).

Net littoral drift is westward and exceeds 300,000 yd^3/yr for the headlands (Research Planning Institute 1983). Transport at Napeague Beach is only 50% of this value (Figure 10), indicating that the headland is the sediment source for Napeague Beach.

No shoreline change information is available for that section of the CBRS unit that includes a portion of the shoreline of Napeague Harbor.

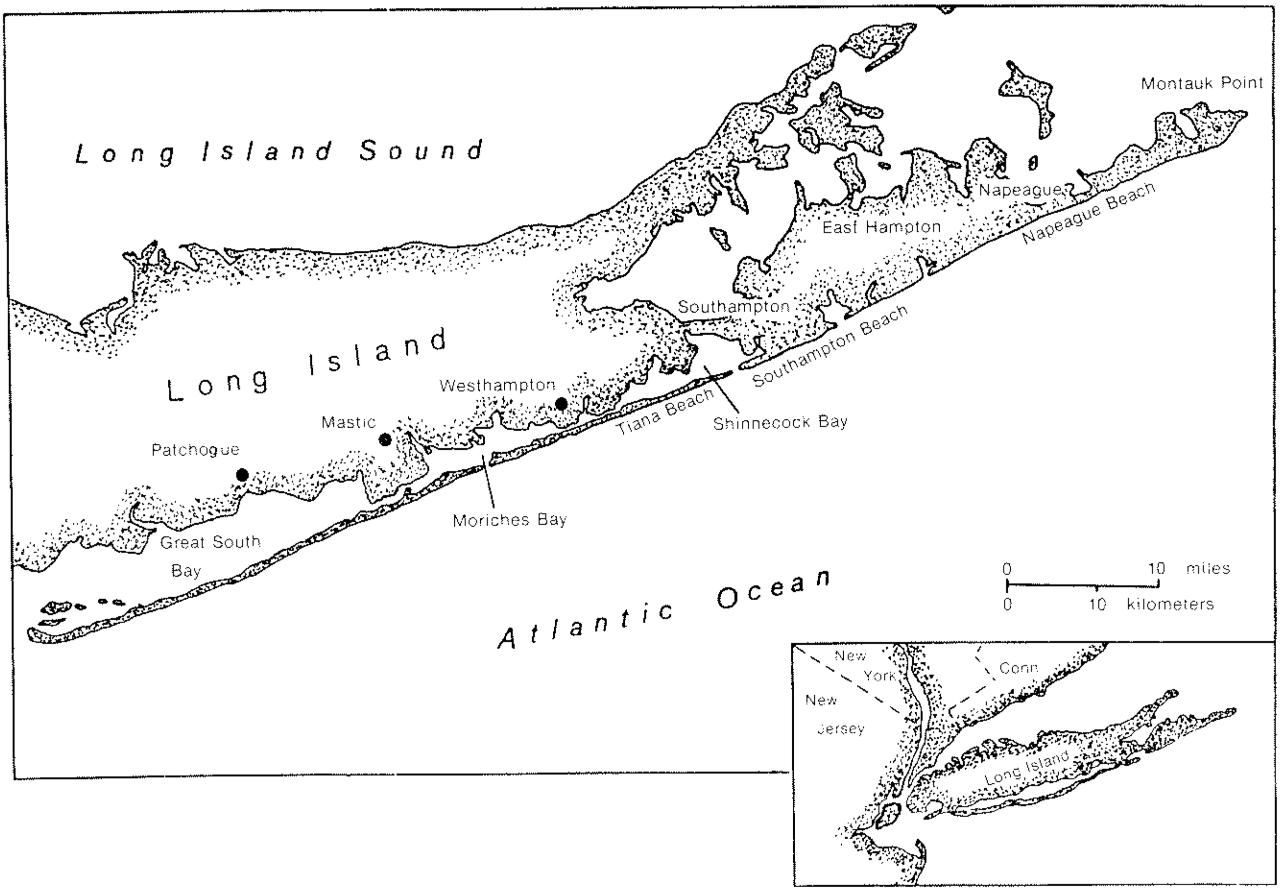

Figure 8. Location of Napeague Beach, Southampton Beach, and Tiana Beach on Long Island, New York (Leatherman and Allen 1985).

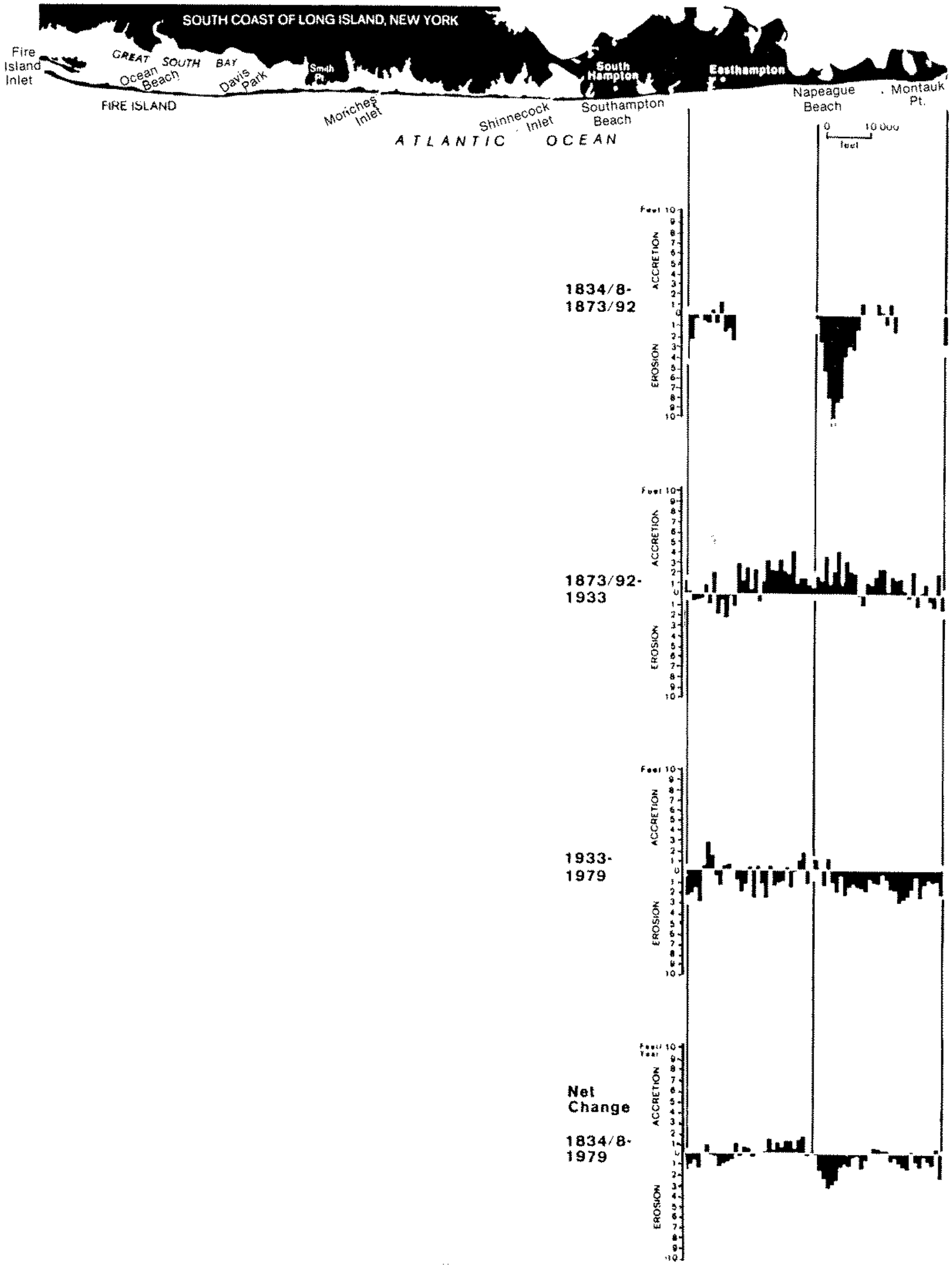

Figure 9. Episodic and net shoreline changes along eastern Long Island, New York, 1834-1979 (Leatherman and Allen 1985).

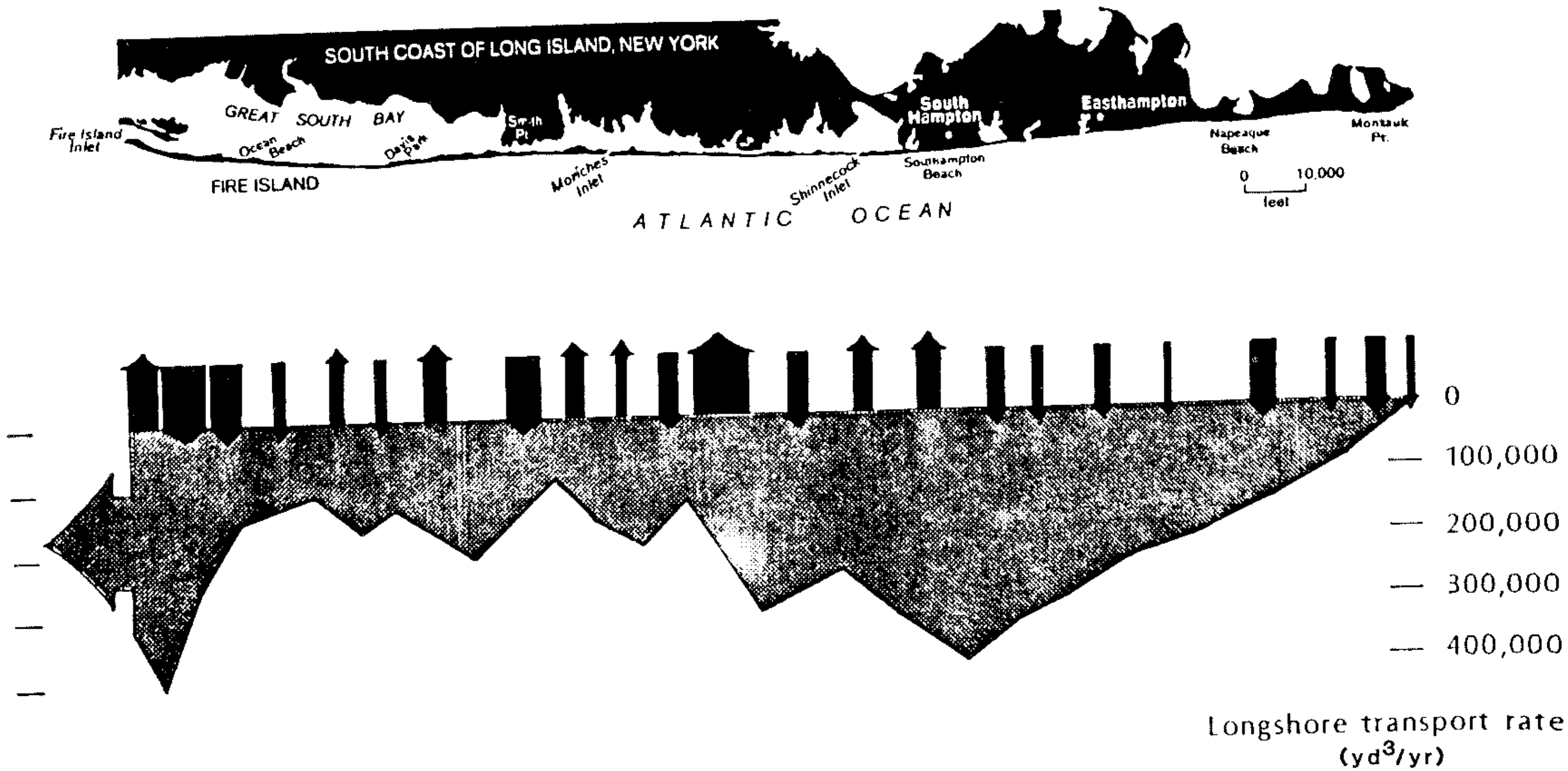

Figure 10. An annualized sediment budget along the south coast of Long Island, New York, 1955-79 (Research Planning Institute 1983).

Coastal Change Processes

Napeague Beach lies within a microtidal environment that has semi-diurnal tides with a mean range of 2.5 ft and a spring range of 3 ft (U.S. Dept. of Commerce 1985). Since Napeague Beach is located at the eastern end of Long Island, it is exposed to dominant waves originating out of the eastern quadrants. Under normal swell conditions, 1.3- to 1.7-ft waves are typical, but storm waves originating from the east and southeast have exceeded 13 ft. Storm surge levels range from 1.5 ft for the worst annual storm to 3.8 ft for 10-year storms (Leatherman and Allen 1985). Major storm activity can be ascribed to hurricanes and extratropical depressions (northeasters). The frequency of severe and unusually severe storms are 9/100 yr and 2.8/100 yr, respectively (Taney 1961).

Although impacts from recent storms, such as those occurring in 1978 (northeaster) and 1985 (Hurricane Gloria), were considered severe, they do not rival those of the unusually severe storms that occurred in 1938 and 1962. The 1938 storm was a September hurricane that made landfall on central Long Island during a rising tide. The result was the creation of major breaches and extensive washover deposits along the barrier shoreline. Inshore wave heights approached 13 ft and storm surge exceeded 8 ft (U.S. Army Corps of Engineers 1963a). The 1962 (Ash Wednesday) storm was a northeaster that hit the coast for five consecutive high tides during a spring tide stage. Storm surge was continuous and exceeded 4.8 ft during the entire interval. The major impacts of these storms were considerable erosion of the shoreline and widespread destruction of shorefront residences. Impacts along the barrier shoreline to the west surpassed those along the headland area where Napeague Beach is situated.

The eroding headland has been designated as one of two principal sediment sources for the Long Island barrier system (Taney 1961; McCormick and Toscano 1981). The effect on Napeague Beach, from the most recent observations, is a locally positive sediment budget that provides some buffering and a natural means of shoreline restoration from the impacts of major storms.

Apparent sea-level rise is an important long-term component of erosional processes. Tide gauge analyses for several stations on, and in close proximity to, Long Island indicate an average sea-level rise of about 1.3 inches per decade (3.2 mm/yr) during the past 40 to 60 years (Hicks 1978).

Management Implications

The eastern end of Long Island has been undergoing persistent erosion although local areas of deposition exist, such as the Napeague Beach area. The marine and aeolian deposits that form Napeague Beach are indicative of net deposition in the past when spit accretion processes dominated and connected adjacent Pleistocene headlands. However, the pattern of shoreline change along Napeague Beach does vary, and management strategies should consider this. The CBRS unit (a small section of Napeague Beach proper) has been undergoing net erosion over the past 150 years, although episodes of accretion have occurred. The factors which govern the shoreline response are poorly understood but are thought to be related to differential resistance to erosion in the headlands with respect to the existing wave climate (McCormick and Toscano 1981).

Development in the vicinity of Napeague Beach has primarily occurred in East Hampton and along the bayside west of Napeague Harbor where the effects of net erosion during historic time have not been observed. Development to the east has been largely related to infrastructure supporting Hither Hills State Park. Less development has taken place within the CBRS unit. Although the unit is wider than many of the barrier islands in the Long Island system, land within the unit is the worst place in the local area to develop. The shoreline here is the narrowest and the lowest, and has been undergoing net retreat over the longest record in historical time. Any development should be directed to the west where shoreline retreat is minor and a sufficiently wide oceanside setback can be delineated.

CBRS UNITS F12--SOUTHAMPTON BEACH AND F13--TIANA BEACH, NEW YORK

Geomorphology

Long Island's barrier island system is moving both longshore and landward. Sediments entering the system are thought to derive primarily from the headlands and offshore sources (Taney 1961: McCormick and Toscano 1981). Net littoral drift from the east has resulted in a rapid longshore extension of the barrier island system. Presently, Tiana and Southampton Beaches (Figure 8) can be described as low-lying, transgressing barriers composed of extensive washover deposits that interface with a 20-ft high linear dune on the oceanside. The bayside of Tiana Beach is underlain by 3 feet of artificial fill material. The eastern terminus of Tiana Beach and the western terminus of Southampton Beach are marked by jetties at Shinnecock Inlet. The CBRS unit only occupies a 1.5 mi section of Tiana Beach starting 3.3 mi west of the inlet. The Southampton Beach CBRS unit begins some 800 ft east of Shinnecock Inlet and extends eastward for about 1.4 mi. The main axes of the CBRS units have been relatively stable, although the width of the islands has decreased. The barrier segment between Shinnecock Inlet and the CBRS units is composed of relict inlet channel deposits which underlie recently deposited washover sands.

Geologic History

The major surficial features along the south shore of Long Island, New York, are Pleistocene and Holocene in origin. The eastern end is a headland composed of remnants of a late Pleistocene moraine deposit. Marine processes associated with the Holocene transgression have since carved these high bluffs or sea cliffs. Materials eroded from the headland are transported to existing barrier spits and barrier islands which are Holocene deposits that lie over a gently sloping Pleistocene outwash plain. Eustatic (worldwide) sea-level rise has largely contributed to a persistent landward retreat of the shoreface from a position at least 100 ft below existing mean sea level. Apparent sea-level rise is currently about 1.3 inches per decade (3.2 mm/yr).

The formation and landward movement of the Long Island barriers have been a subject of controversy over the past decade. One viewpoint suggests that the present barrier system formed after an ancestral barrier chain was overstepped or drowned in-place as a result of a rapid rise in sea level (Rampino and Sanders 1981). The present barrier system is inferred to be a Pleistocene ridge capped by reworked Holocene sediments. An alternative hypothesis suggests that the barrier islands have retreated continuously as an equilibrium response to eustatic rise in sea level and the availability of sediment (Swift and Moslow 1982). A reevaluation of the original cores that formed the basis of the drowning theory and an interpretation of newly acquired geophysical data indicate that continuous shoreface retreat seems to be the principal mode of migration for the current barrier system (Leatherman and Allen 1985).

Modern Changes

A 145-year record of shoreline change exists for the Tiana and Southampton Beach area (Figure 11). Records between 1834 and 1873 indicate that the island extending westward from Sedge Island (within the Tiana Beach CBRS unit) incurred net erosion with retreat averaging 4.3 ft/yr. The eastern end was accretional in the vicinity of the CBRS unit, with a maximum accretion rate of 5.3 ft/yr. However, this was flanked by another shoreline segment undergoing erosion at a rate of 3.3 ft/yr. The interval between 1873 and 1933 was one of marked stability and net accretion throughout all but the Sedge Island and Lanes Island (between Shinnecock Inlet and the Tiana Beach CBRS unit) areas. A maximum accretion rate of 4.3 ft/yr occurred to the east of the Tiana Beach unit. The last interval, between 1933 and 1979, was a major episode of erosion. The 1938 hurricane, the 1962 (Ash Wednesday) storm, inlet creation, inlet migration, and jetty construction were major factors contributing to net erosion along Tiana Beach. With jetty construction, net deposition occurred updrift of the inlet at Southampton Beach (CBRS unit F12); correspondingly, net erosion resulted downdrift at Tiana Beach (CBRS Unit F13). An average shoreline accretion rate of 11 ft/yr occurred updrift at Southampton Beach, but the Tiana Beach shoreline retreated at a rate of 8.5 ft/yr. The net accretion during the previous century was nullified with net shoreline change for the entire period approaching 1.6 ft/yr of erosion.

Coastal Change Processes

The Southampton-Tiana Beach barrier islands are situated within a microtidal environment which has a 3.1 ft mean range and a 3.8 ft spring range (U.S. Dept. of Commerce 1985). Locally generated waves originate from the southwest, but the principal forcing occurs from an unrestricted fetch extending from the northeast to the south. Annual sediment transport at Shinnecock Inlet approximates 300,000 yd^3/yr (Panuzio 1968; Research Planning Institute 1983).

Disturbances such as hurricanes and extratropical storms (northeasters) affect the coast. Under normal swell conditions, 1.3 to 2.6 ft waves are typical, but storm-generated waves have exceeded 13 ft. Storm surge level ranges from 1.5 ft for the worst annual storm to 3.8 ft for storms with a 10-year

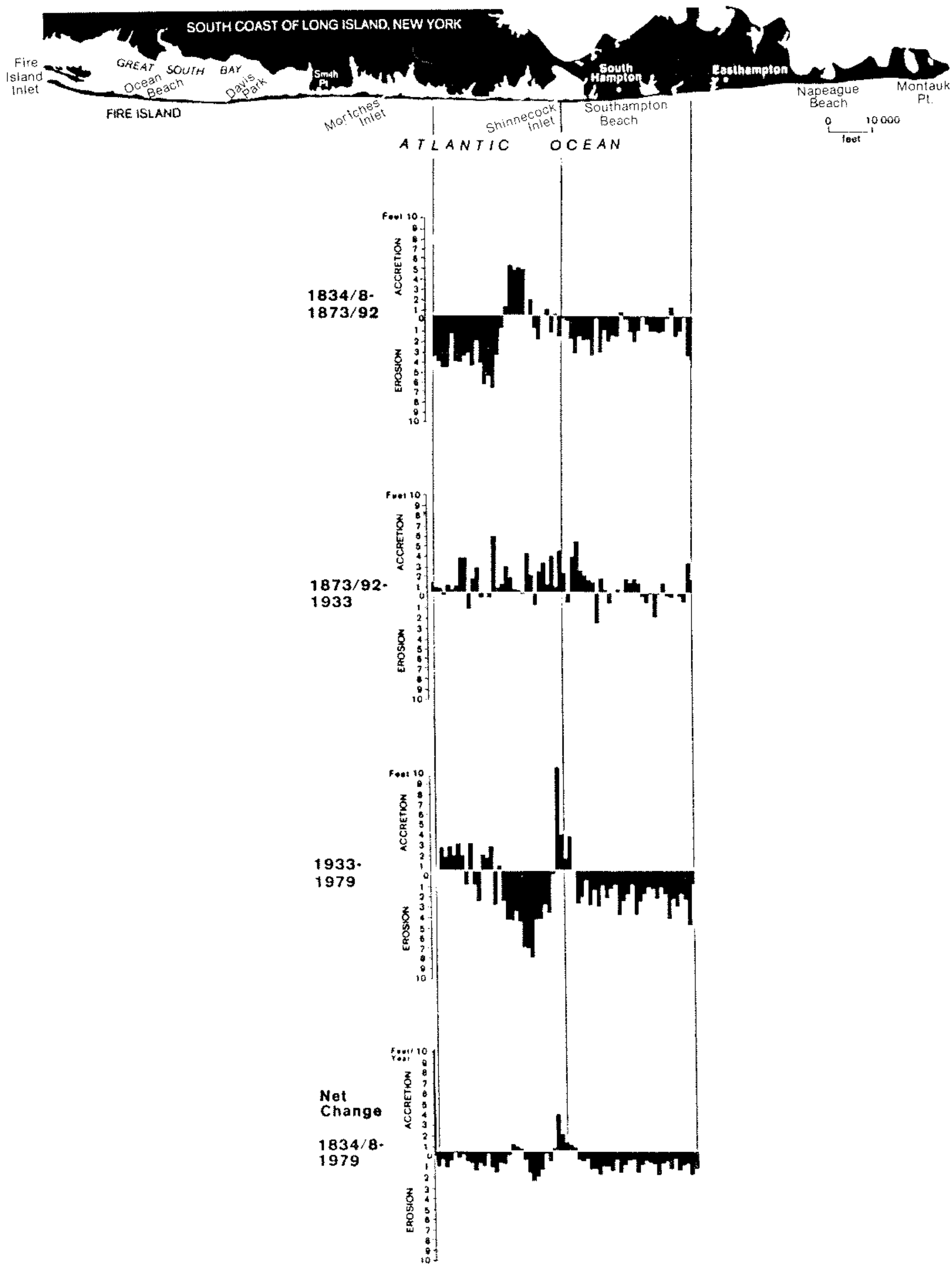

Figure 11. Episodic and net shoreline change along the south-central coast of Long Island, New York, 1834-1979 (Leatherman and Allen 1985).

recurrence interval (Leatherman and Allen 1985). Storm frequencies on a 100-year interval for severe and unusually severe storms are 9 and 2.8, respectively (Taney 1961). One significant hurricane impact in the area was the creation of Shinnecock Inlet in 1938. The frequency of northeasters exceeds hurricanes. These storms result in severe beach erosion, widespread washovers, and destruction of shorefront property and infrastructure.

Management Implications

The CBRS units are situated within particularly sensitive segments of the south shore barrier island complex. The Tiana Beach area has experienced net erosion during the last 50 years. The recent construction of jetties between 1952 and 1954 at Shinnecock Inlet has induced a landward offset of the Westhampton-Tiana Beach barrier as a result of sediments being impounded updrift at Southampton Beach and diverted offshore to the lee of the structures. The effect west of Shinnecock Inlet has been a decreasing island width and increasing susceptibility to flooding.

The CBRS units are vulnerable to overwash processes. The 1938 hurricane and the 1962 storm-generated washovers that extended to the back-barrier marsh over major sections of both units and completely to the bayshore at narrow segments of the beaches.

The above effects strongly imply that the Tiana Beach unit is not a good site for development. Stability of the barrier shoreline downdrift of the inlet would be difficult to achieve. An integrated approach of inlet sediment bypassing, beach nourishment, and artificial dune construction might help; however, its implementation in such a narrow and relatively low-lying barrier would require considerable maintenance and continual funding. The Southampton Beach unit is also a narrow and relatively low-lying barrier but it is undergoing accretion due to the jetties at Shinnecock Inlet, largely at the expense of Tiana Beach.

The bays in this general area serve as important nursery grounds for finfish and the bay bottoms and tidal flats support benthic organisms of commercial and recreational importance. The particular status of the bottom habitat within the part of Shinnecock Bay that lies within the CBRS is unknown. In the case of the Tiana Beach unit, the bay bottom is the site of spoil deposition.

CBRS UNIT L03A--SHACKLEFORD BANKS, NORTH CAROLINA

Geomorphology

Shackleford Banks is the southernmost extent of a nearly continuous chain of barrier islands along the North Carolina coast, collectively referred to as the Outer Banks (Figure 12). The island is approximately 9 mi long and 0.25 to 1.0 mi wide. It is oriented in a northwest-southeast direction, in contrast to the northeast-southwest orientation for most of the other Outer Banks barrier islands. Maximum elevations on Shackleford Banks are 35 ft above mean sea level (Susman and Heron 1979).

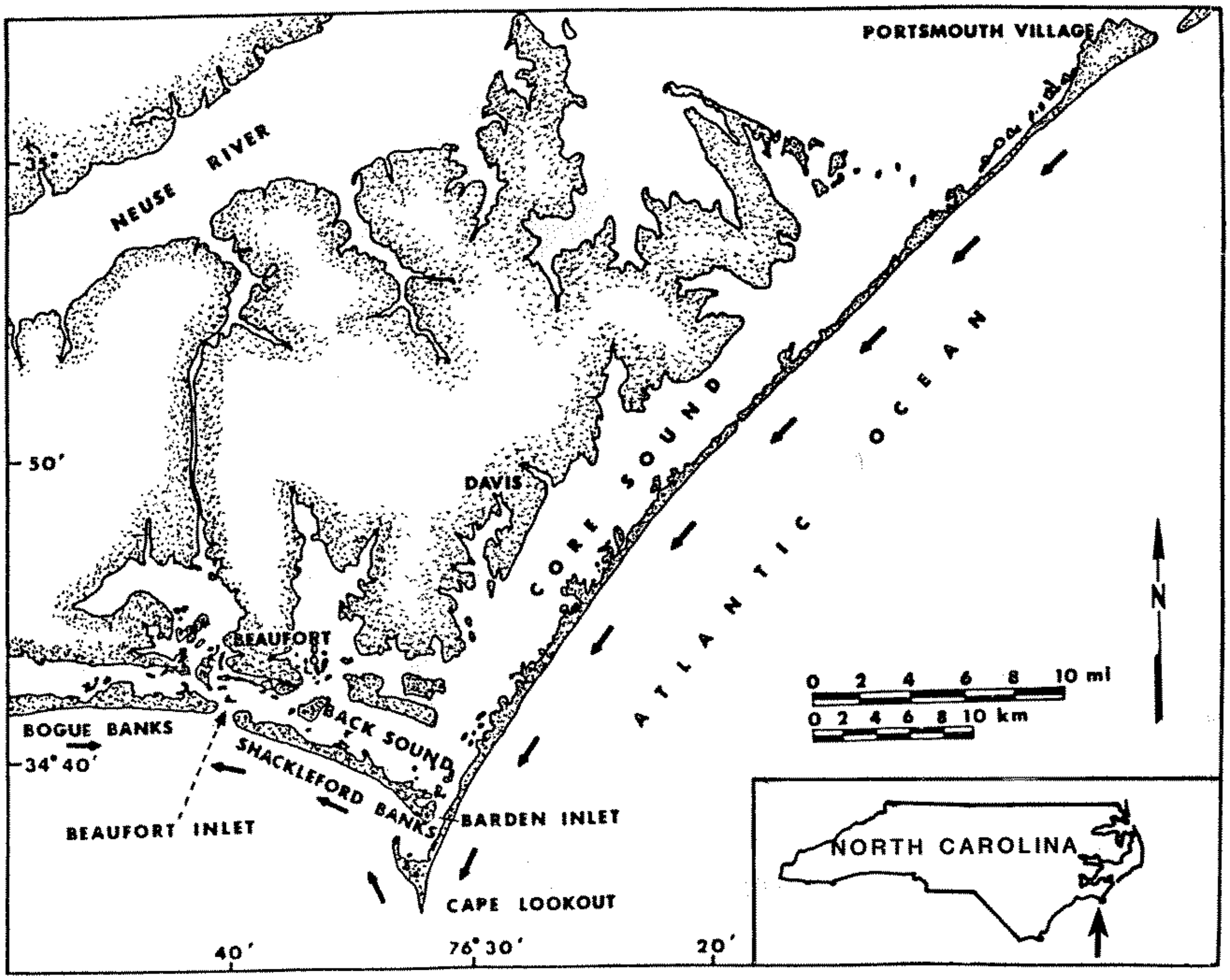

Figure 12. Location of Shackleford Banks, North Carolina. Arrows indicate the direction of net longshore sediment transport (Hamilton 1977).

Shackleford Banks is bordered on the southeast by Barden Inlet, which separates the island from Core Banks and Cape Lookout, and on the northwest by Beaufort Inlet, separating Shackleford from Bogue Banks (Figure 12). To the north, Back Sound, a relatively shallow, narrow, microtidal (less than 6 ft tidal range) lagoon, separates Shackleford from a geologically unrelated Pleistocene mainland, which is part of the North Carolina Coastal Plain. Lookout Bight and the Atlantic Ocean border the island to the south (Figure 13).

The relatively long, linear, and narrow morphology of Shackleford Banks is characteristic of a microtidal barrier island. The most prevalent individual morphologic feature on the island is a series of recurved beach ridges (Figure 13). These ridges are present along the entire length of the island and are thought to have formed by spit accretion associated with lateral tidal inlet migration (Fisher 1967; Susman and Heron 1979). Beach ridge orientation suggests a westerly inlet migration.

Variations in morphology along Shackleford Banks are best illustrated by a series of shore-perpendicular transects developed by Brauer (1974) (Figure 14). The western tip of the island is dominated by a wide beach, broad intertidal flats, and incipient dunes (transect A-A', Figure 14). This morphology was formed by spit accretion due to migration of Beaufort Inlet. The remainder of the western half of the island is characterized by a well developed foredune ridge, recurved dune ridges, vegetated interdune swales, and a climax maritime forest (transect B-B', Figure 14). The eastern half of the island is dominated by arcuate storm overwash fans, vegetated barrier flats, incipient dunes, and broad expanses of salt marsh (transects C-C' and D-D', Figure 14). On the landward side of the barrier, large, lobe-shaped features, capped by salt marsh and extending out into Back Sound, have been interpreted as relict flood-tidal deltas (Figure 13).

Geologic History

The Holocene geologic history of Shackleford Banks has been examined in detail through a series of deep borings by Susman and Heron (1979) and Heron et al. (1984). The island's history has been interpreted as being dominated by large scale erosion from tidal inlet migration and subsequent infilling. A stratigraphic cross-section of the island shows that most of the Holocene subsurface is a thick, laterally extensive sequence of tidal inlet-fill deposits. At least two, and possibly several, tidal inlets have been active during the island's geologic history (Figure 13). As a result of lateral inlet migration, the sedimentary record of Shackleford Banks has been totally reworked. It is likely however, that in response to rising eustatic sea level during the late Holocene, the island has migrated landward from a much further seaward position.

Modern Changes

The most significant events that have prevailed on Shackleford Banks during its "recent" (last 250 years) history are (1) beach erosion along the eastern half of the island; (2) lateral spit accretion due to tidal inlet migration at

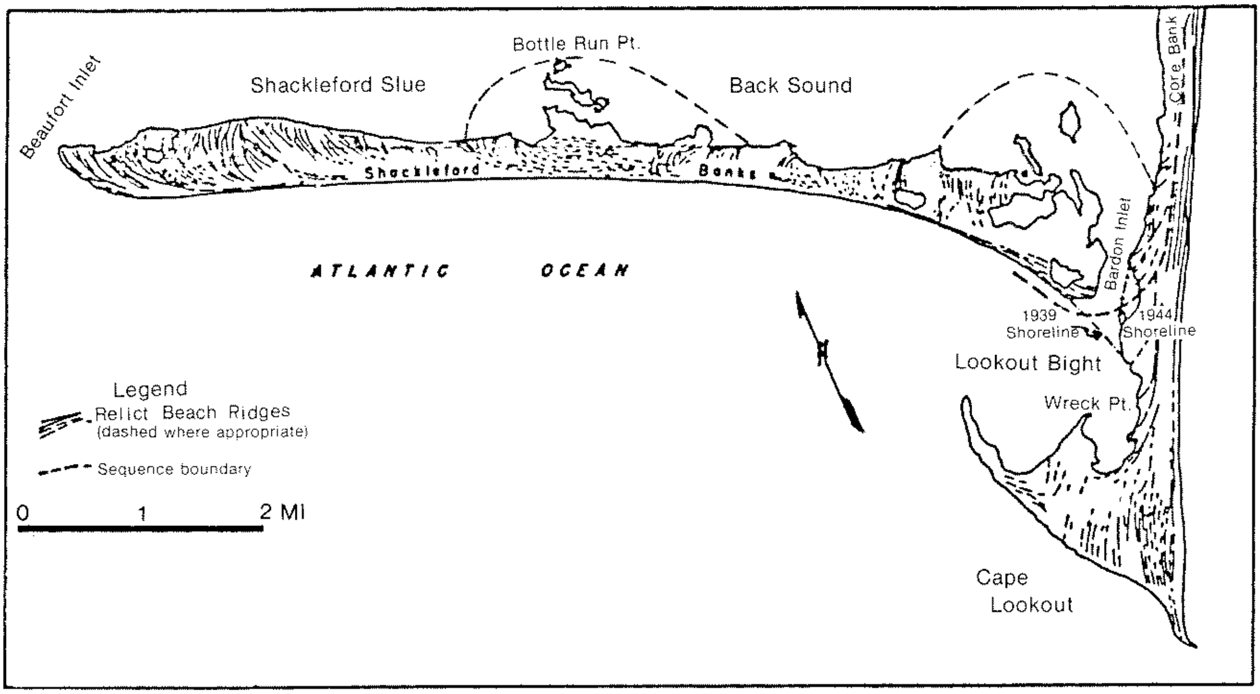

Figure 13. Generalized orientation of beach ridges on Shackleford Banks, North Carolina. The dashed lines forming an arcuate pattern on the soundside of the island enclose relict flood-tidal deltas (modified from Fisher 1967).

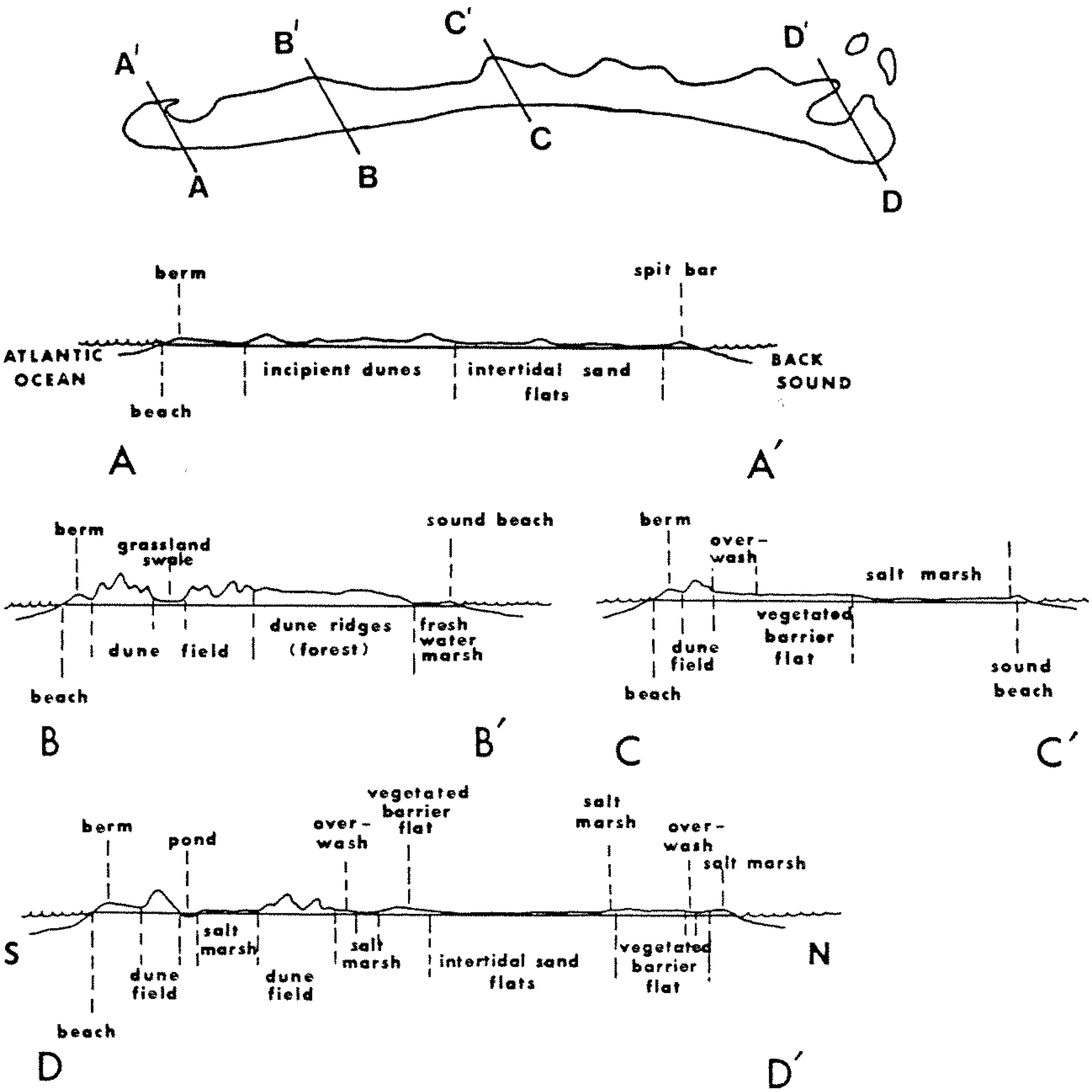

Figure 14. Surface morphology of four cross-sections of Shackleford Banks, North Carolina (Brauer 1974).

the western end of the island; and (3) overwash deposition and marsh accretion on the landward margin (sound side) of the island. The combination of these events has resulted in a northwestern migration of Shackleford Banks, having both lateral (alongshore) and landward components of movement. However, the lateral component of this movement has slowed considerably due to rock jetties that are controlling channel migration in Beaufort Inlet and subsequent spit accretion at the west end of Shackleford.

An analysis of historic maps and charts dating back to 1585 shows that at least two, and possibly three, tidal inlets have been active on Shackleford Banks (Fisher 1962). Beaufort Inlet has been open since at least 1585, and has migrated laterally at least 0.8 mi since 1939 to its present position. Barden Inlet was open during the late 18th or early 19th centuries, but was closed by natural processes around 1850 (Fisher 1962). Barden Inlet was reopened by an unnamed hurricane in 1934 and remains open today only by the active dredging of its main channel.

The beaches on the eastern half of Shackleford Banks have been eroding since 1953, probably as a result of the dredging at Barden Inlet (Brauer 1974). Brauer (1974) suggested that the seaward margin of the eastern half of the island is sediment starved. Erosion and accretion trends for the beaches of Shackleford Banks from 1939 to 1964 have been compiled by Brauer (1974). On the northern (soundside) shore of Shackleford, erosion has been more extensive than accretion. Erosion has occurred since 1939 along most of the sound side beaches except where overwash fans capped by salt marsh have prograded out into the lagoon.

Coastal Change Processes

The major agents of coastal change at Shackleford Banks, as well as at all of the Outer Banks of North Carolina, are waves, tides, and storms. Mean tidal range in this area varies from 1.0 to 3.5 ft. Although the tidal range is relatively low, tidal processes should not be discounted in the Cape Lookout area. Tidal currents have been measured in excess of 3.8 ft/s (Sarle 1977), and are responsible for transport and deposition of fine- and coarse-grained sediment especially in the tidal inlets, tidal deltas, and lagoons. The size of the tidal prism is an important factor in controlling the magnitude and duration of tidal currents, and in determining flood or ebb dominance of sediment transport in tidal inlets and backbarrier environments (Moslow and Tye 1985).

Although tides and tidal prism are important, the most significant coastal processes affecting Holocene sedimentation in the area are waves and storms. Mean annual wave height is 5.6 ft with wave heights exceeding 6.5 ft approximately 30% of the year (Nummedal et al. 1977). These values are among the highest for the U.S. east coast and are primarily responsible for the wave-dominated barrier shoreline morphology.

Northeasters and hurricanes also have an important impact on sedimentation patterns in the area. Storm-related processes redistribute coarse-grained sediment on the ocean margin of tidal inlets, tidal deltas, and lagoons.

Large volumes of sand are transported across low-lying barriers during storms and deposited in the form of washover fans. The North Carolina coast has a history of 150 recorded hurricanes since 1585. An average of 1.64 hurricanes per year affect the Shackleford area (Crutcher and Quayle 1974) with recorded storm surges up to 7.5 ft above mean sea level (U.S. Army Corps of Engineers 1976).

Management Implications

Shackleford Banks is a relatively unstable barrier. Geologic and historic evidence indicates a long history of island breaching and storm overwash, inlet formation, inlet migration, and inlet closure. These active processes and others help to maintain a high diversity of fish and wildlife habitat. For example, the prevalent formation of overwash fans on the eastern half of the island during high energy storms, provides a low relief infilling of the soundside of the island which facilitates the development of salt marsh habitat. Lateral inlet migration and associated spit accretion along with aeolian processes have resulted in well-developed dune ridges supporting a climax maritime forest along the central and western portions of the island.

These same geologic processes, however, are obviously quite destructive to cultural developments. There was a thriving whaling village on the island in the 18th and early 19th century, but all permanent residents left the island after a devastating hurricane in the late 19th century destroyed the village. The general inaccessibility of Shackleford Banks has helped to keep it largely undeveloped.

The rock jetty at the west end of Shackleford has affected the westward migration of the island. Dredging operations in Barden Inlet have exacerbated erosion rates on the eastern end of the island. Without dredging, Barden Inlet would close and Shackleford Banks would become attached to Core Banks. Past history indicates that Barden Inlet would probably reopen intermittently as a result of major storms. The maintenance of the inlet does result in some sediment deprivation to the eastern end of the island. More innovative use of the dredge spoil (i.e., as beach nourishment material) that results from Barden Inlet maintenance would help to reduce erosion rates on the eastern half of the island.

This CBRS unit is now a part of the National Seashore System and the Department of the Interior has recommended that it be dropped from the CBRS if otherwise protected areas are to remain outside of the CBRS.

CBRS UNIT M09--EDISTO COMPLEX, SOUTH CAROLINA

Geomorphology

The Edisto Complex is located on the central South Carolina coast and comprises the northern one-half to two-thirds of what is commonly referred to as Edisto Island. The part of Edisto Island that is in the CBRS unit includes

Edingsville Beach to the south and Botany Bay Island to the north (Figure 15). Edingsville Beach is approximately 2.0 mi in length and is bounded by Jeremy Inlet to the south and Frampton Inlet to the north. Botany Bay Island is approximately 3.0 mi in length and extends alongshore from Frampton Inlet to South Creek.

Edingsville Beach and Botany Bay Island are very similar morphologically. Both are Holocene transgressive barrier islands separated from the Pleistocene mainland by a 0.6 to 1.8 mi wide salt marsh. The southernmost part of Edingsville Beach is attached to Pleistocene beach ridges and may be geologically defined as a transgressive mainland beach. Both Edingsville Beach and Botany Bay Island are characterized by narrow sand and shell beaches that are attached to a broad salt marsh extending inland. Exposed rooted salt marsh muds and oyster shell debris are common on both beaches. Overwash fans and terraces overlie salt marsh along the seaward margin of the barriers. A few meandering tidal creeks and ephemeral tidal inlets (Jeremy and Frampton inlets are examples) are also present. On their landward sides, both islands contain a small number of shore-oblique beach ridges whose morphology is suggestive of lateral (northeastern) spit growth (Figure 16). Although much wider at one time in the recent past, both Edingsville Beach and Botany Bay Island are presently very narrow, averaging only 0.6 to 1.2 mi in width.

Geologic History

A recent geologic history for Edisto Island, including Edingsville Beach and Botany Bay Island, was reconstructed by McCants (1982), principally from morphologic evidence. The difference in orientation of the present-day beach ridges compared to the present shoreline (Figure 16), combined with the identification of paleo-floodplains along the Edisto River from Landsat imagery, led McCants to develop the following paleogeographic reconstruction for Edisto Island (Figure 17). Edisto Island retreated landward at moderate rates, from prior to 5,000 years B.P. to about 4,200 years B.P., due to sea-level rise. Following a rapid rise in relative sea level about 4,200 years B.P., the South Edisto River captured most of the flow (and sediment load) of the North Edisto River. The reduced sediment supply combined with a higher stand of sea level initiated a phase of increased shoreline along the island around 4,200 years B.P. With the stabilization of sea level about 3,000 years B.P., shoreline erosion was halted. During the past 3,000 years, the shoreline of Edisto Island has experienced net accretion although erosion has dominated since at least 1859 (McCants 1982).

Modern Changes

The recent history of the Edisto Complex has been dominated by rapid shoreline erosion. From analysis of the earliest reliable charts of the South Carolina coast (1853), Hayes et al. (1979b) determined that the Botany Bay Island shoreline (all within CBRS) has eroded over 3,280 ft. This rapid erosion has been due principally to a decrease in the natural sediment supply from longshore transport and a focusing of wave energy on the northeastern part of the Edisto Complex caused by wave refraction around the North Edisto Inlet ebb-tidal delta. This massive erosion from waves and sediment deficiency

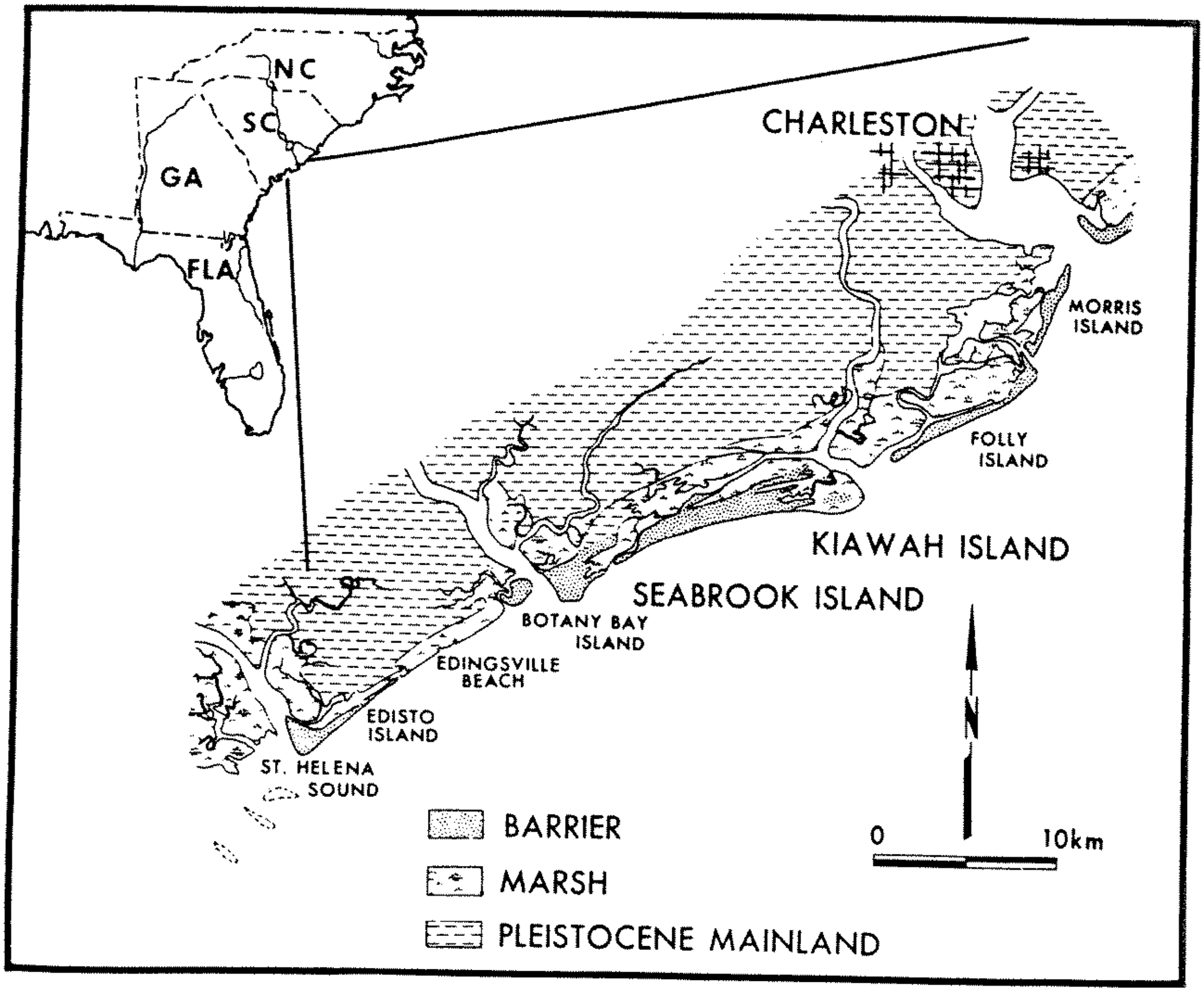

Figure 15. Location of Edingsville Beach and Botany Bay Island, South Carolina (Moslow 1980).

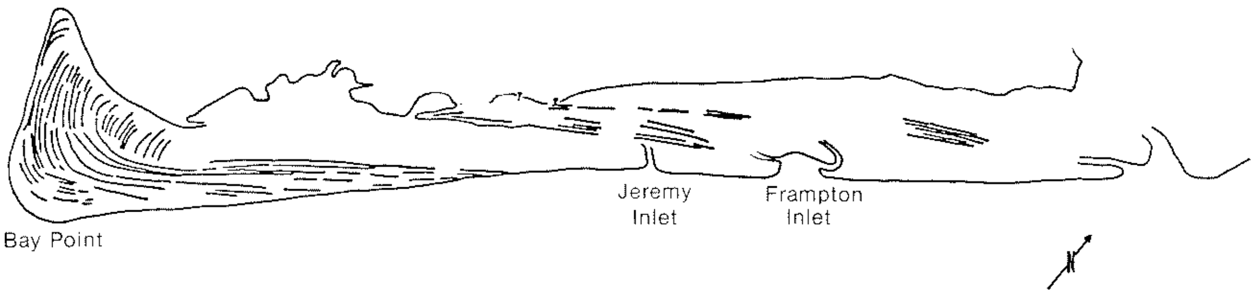

Figure 16. Beach ridge tracings for Edisto Island, Edingsville Beach, and Botany Bay Island (McCants 1982).

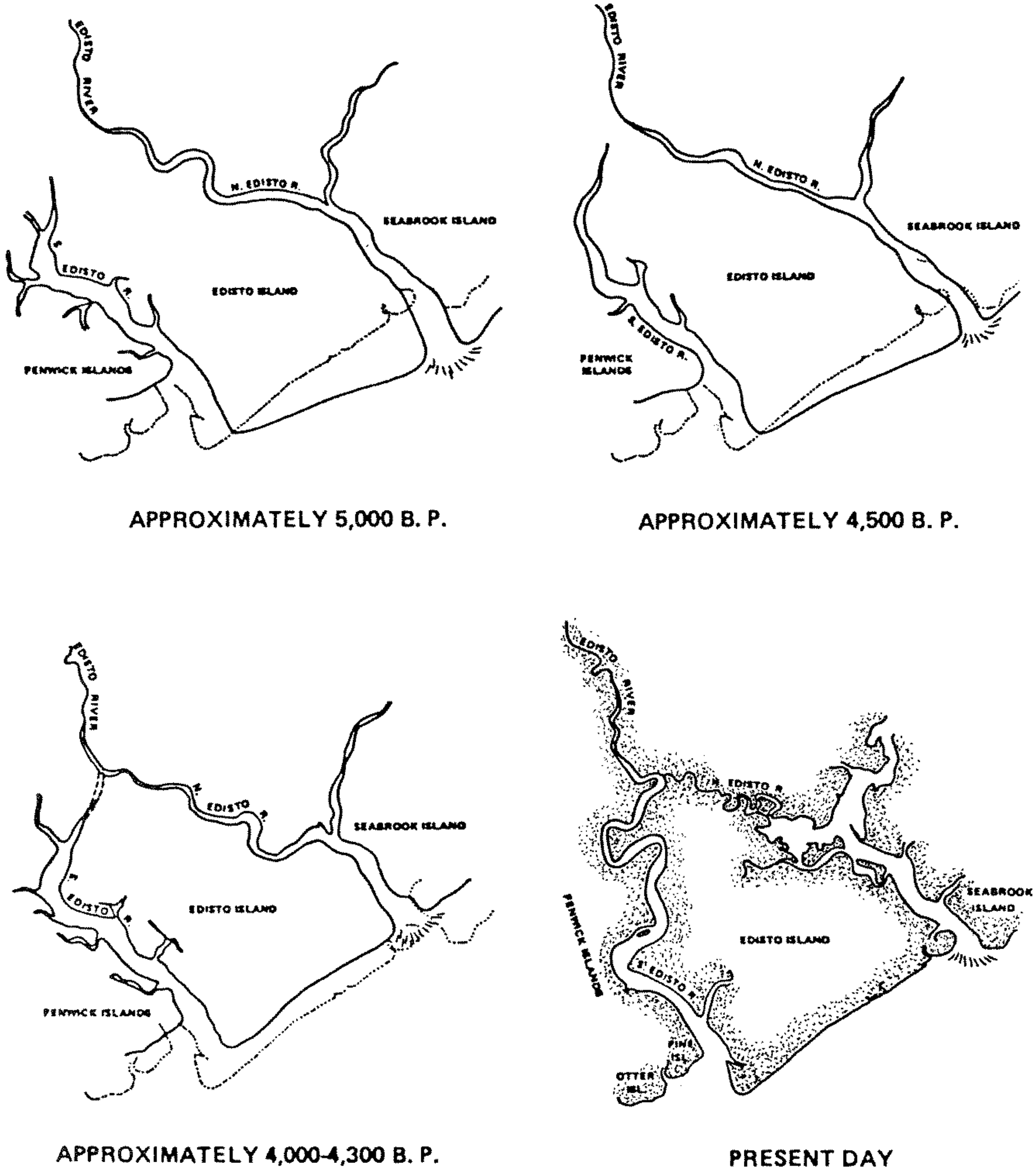

Figure 17. Paleogeographic reconstruction of the Holocene Edisto Island complex from 5,000 years B.P. to the present (McCants 1982).

along portions of the coastline is part of a natural cycle that is discussed in the next section.

An analysis of historical shoreline changes on the Edisto Complex was conducted by Hulse and Kanes (1972), Hayes et al. (1979b), and McCants (1982). The following changes in shoreline position and orientation have been observed.

1. Botany Bay Island and the entire Edisto Complex have continually retreated since 1853;

2. The shoreline at Botany Bay Island eroded 643 ft from 1852 to 1924;

3. The amount of shoreline erosion and landward retreat continuously increases in a northeasterly direction from a null point southwest of Jeremy Inlet (southwest boundary of the CBRS unit), producing a substantial change in shore configuration;

4. Tidal creeks have been intersected by shoreline retreat to form tidal inlets with an appreciable ebb-tidal delta;

5. The most rapidly changing areas of the Edisto Complex are at the mouths of Frampton (within the CBRS unit) and Jeremy Inlets. Frampton Inlet in particular has migrated over 980 ft alongshore from 1963 to 1972 (Hulse and Kanes 1972; Figure 18).

<u>Coastal Change Processes</u>

The Edisto Complex is influenced by prevailing winds from the south-southwest and storm winds from the northeast. The northeast winds, though less frequent, are strongest, causing waves and longshore sediment transport to be directed southeast along the Edisto Complex.

Mean tidal range in the Edisto Complex is approximately 5.2 ft with spring tides ranging up to 6.1 ft. These tides generate strong currents that are responsible for transport of sediment in tidal creeks and inlets. This relatively high tidal range results in exposure of wide portions of the beach to wave action. Beach erosion is most likely during storms that coincide with spring tides (Hayes et al. 1979b).

Storm surges, produced by strong winds pushing water landward and increasing water levels, can result in significant erosion and flooding. The highest storm surges recorded in South Carolina are associated with hurricanes. The maximum storm surge observed on the South Carolina coast was 15.5 ft during the passage of Hurricane Hazel in 1954 (Myers 1975). Hurricane force winds (exceeding 75 mph) have a frequency of occurrence of once every 14 years (Myers 1975).

A field study conducted during the summer of 1978 on Seabrook Island, immediately adjacent to and updrift from the Edisto Complex, measured average

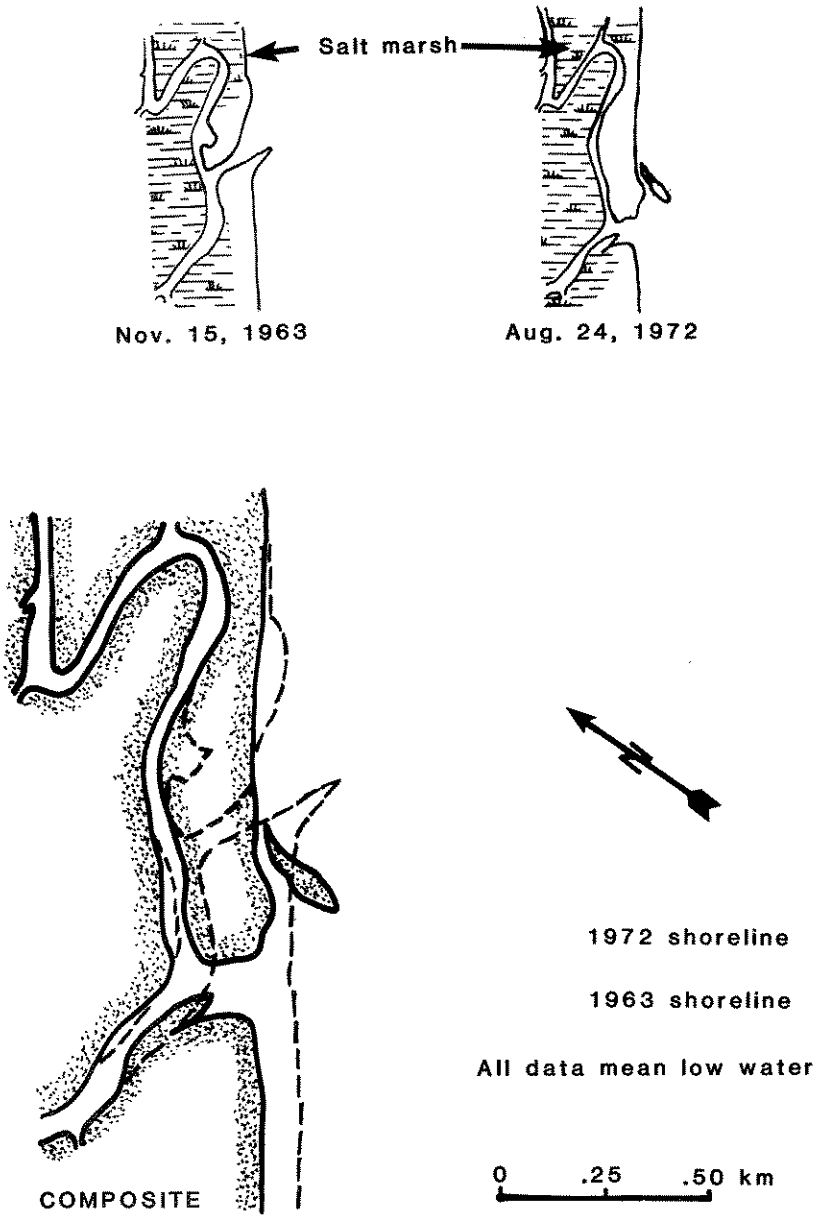

Figure 18. Shoreline changes around Frampton Inlet, 1963-72 (modified from Hulse and Kanes 1972).

wave heights along the shoreline of 0.3 to 1.5 ft (Hayes et al. 1979b). The wide range in observed wave heights was a function of the direction of wave approach from the east-southeast and the positioning of offshore shoals, especially Deveaux Bank. These intertidal shoals (Figure 19) tend to dampen wave energy that would normally break on the shoreline. Also, the refraction of waves over the shoals can serve to focus wave energy on various portions of adjacent barrier islands. Much of the observed shoreline erosion on the Edisto Complex is believed to be due to wave attenuation over the North Edisto Inlet ebb-tidal delta shoals, including Deveaux Bank. The inlet and shoals have also served as a barrier to longshore sediment transport along the Edisto Complex. As a result of this process, sand remains trapped on the updrift (Seabrook Island) side of North Edisto Inlet, accounting for the landward offset and erosion of the sediment-starved Edisto Complex on the downdrift side (Hayes et al. 1979b; Figure 19).

Management Implications

The analysis of historical maps and charts from 1851 to the present has documented long-term trends of shoreline instability and erosion for almost all of the Edisto Complex. This trend is due principally to natural processes of waves, tides, wave refraction and longshore sediment transport. For the short term (next decade) this trend of erosion is likely to continue. Thus, construction or development of any kind along the Edisto Complex should be discouraged.

The areas of most rapid change are at the mouths of Frampton and Jeremy Inlets and the area immediately downdrift of North Edisto Inlet along Botany Bay Island. These areas will need to be monitored on a regular basis to update shoreline position and configuration. The remainder of the Edisto Complex shoreline can be expected to erode at an equally continuous but generally slower rate.

CBRS UNIT M12--ST. PHILLIPS ISLAND COMPLEX, SOUTH CAROLINA

Geomorphology

The St. Phillips Island Complex is composed of four barrier islands along the southern South Carolina coast. The islands lie between Port Royal Sound to the southwest and Fripps Island to the northeast (Figure 20). These four islands (Bay Point, St. Phillips, Capers, and Pritchards) are all entirely within the CBRS unit (Figure 21). They vary markedly in terms of geomorphology and depositional history and have been the site of very few specific geologic investigations. However, despite these latter limitations, some general geologic observations can be made from available maps, charts, and air photographs.

The St. Phillips Complex can be divided into northeastern and southwestern halves, each consisting of two barrier islands, separated by the relatively deep (more than 30 ft) Trenchards Inlet. All four islands are rather short and wide, and are separated from each other by stable tidal inlets. These islands average 1.0 to 2.0 mi in length and 1.0 to 3.0 mi in width. All of the islands have narrow, erosional beaches that are backed by broad expanses of salt marsh and meandering tidal creeks. The morphology and shape of these barrier islands is typical of those found in a mesotidal (6-13 ft tidal range) shoreline environment. Specific morphologic features are as follows.

Bay Point Island. The southernmost of the four barriers in the St. Phillips Complex, Bay Point Island, is located at the mouth of Port Royal Sound. Geologically, Bay Point Island is an extension of St. Phillips Island to the north. Major morphologic features include broad expanses of salt marsh and several meandering tidal creeks. Broad sandy tidal flats front the island on its western and southern shores, which border Port Royal Sound and the Atlantic Ocean. A small recurved spit and accretional beach at the southwestern tip of the island (Bay Point) are associated with the landward attachment of a large, shore-perpendicular tidal sand ridge.

St. Phillips Island. The southern half of St. Phillips is composed of a series of shore-parallel beach ridges. The intervening topographic lows, or swales, contain salt marshes. The northern part of the island is a network of tightly meandering tidal creeks that dissect a broad expanse of salt marsh.

Capers Island. Capers Island is a transgressive barrier island whose morphology is dominated by the activity of adjacent tidal inlets (Trenchards Inlet to the south, Pritchards Inlet to the north). Almost all of Capers Island is covered by salt marsh that is dissected by several meandering tidal creeks. The beach is erosional, relatively narrow, and backed by a series of arcuate washover fans and terraces (Hayes et al. 1979a).

Pritchards Island. The northernmost of the four barriers comprising the St. Phillips Complex, Pritchards Island is one of several beach-ridge barrier islands found along the South Carolina-Georgia border. Sandy, shore-parallel beach ridges are found along the entire length of island. Salt marsh and meandering tidal creeks are dominant features landward of the beach ridges, in low areas between beach ridges, and in the north-eastern (updrift) part of the island bordering Skull Inlet.

Geologic History

The geologic history of any shoreline area can be best determined from subsurface core or borehole data. Since no known studies with this type of data base exist for the St. Phillips Island Complex, its geologic history must be inferred from barrier island geomorphology. The presence of a series of shore-parallel beach ridges on St. Phillips and Pritchards Islands suggests a history of long-term (3,000 to 4,000 years) seaward progradation. Capers and Bay Point Islands lack any significant depositional features such as beach ridges and are examples of transgressive barriers that have been dominated by a history of long-term erosion and landward shoreline retreat (Brown 1976).

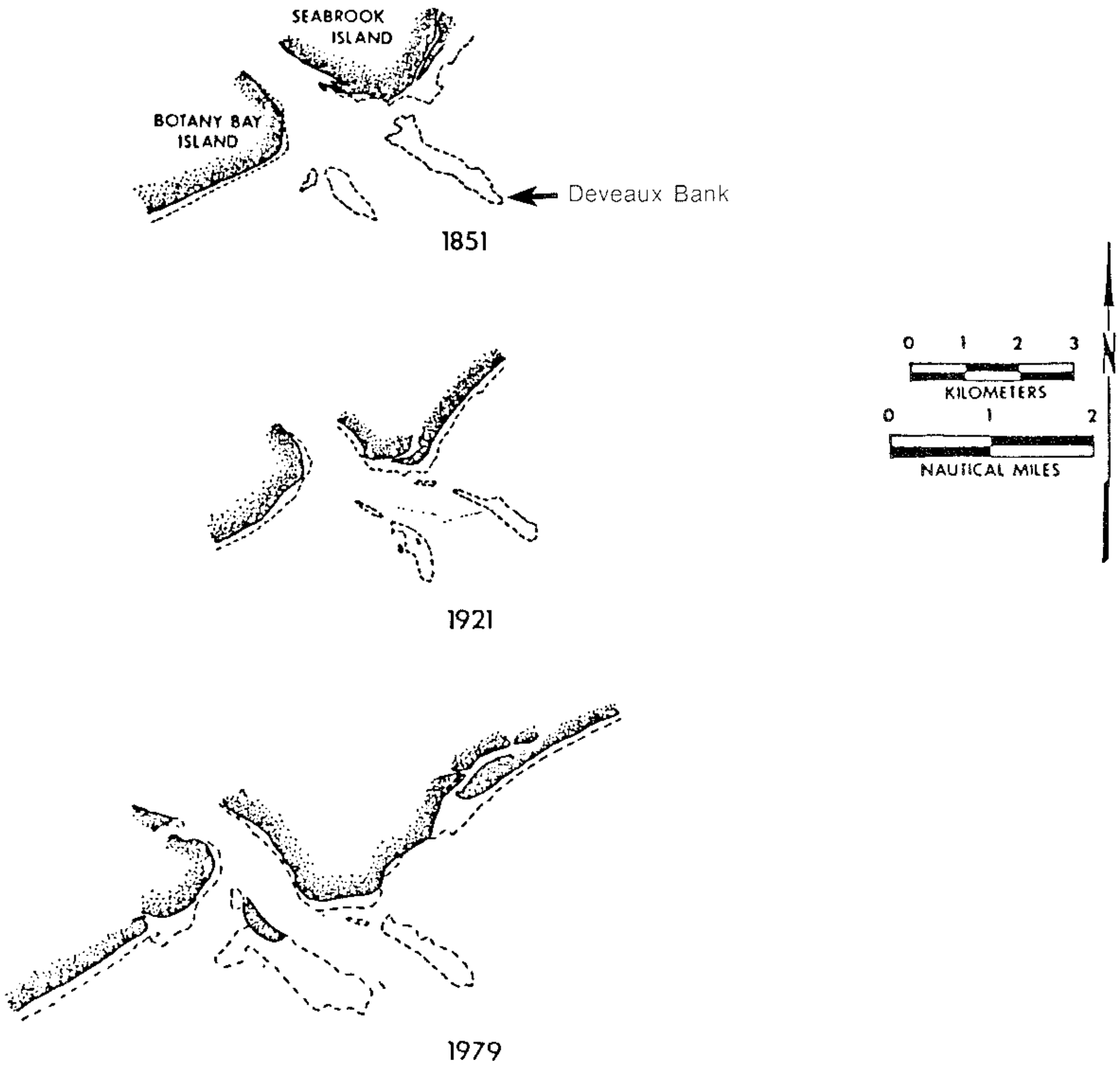

Figure 19. Shoreline changes around North Edisto Inlet, 1851-1979 (Zarillo et al. 1984).

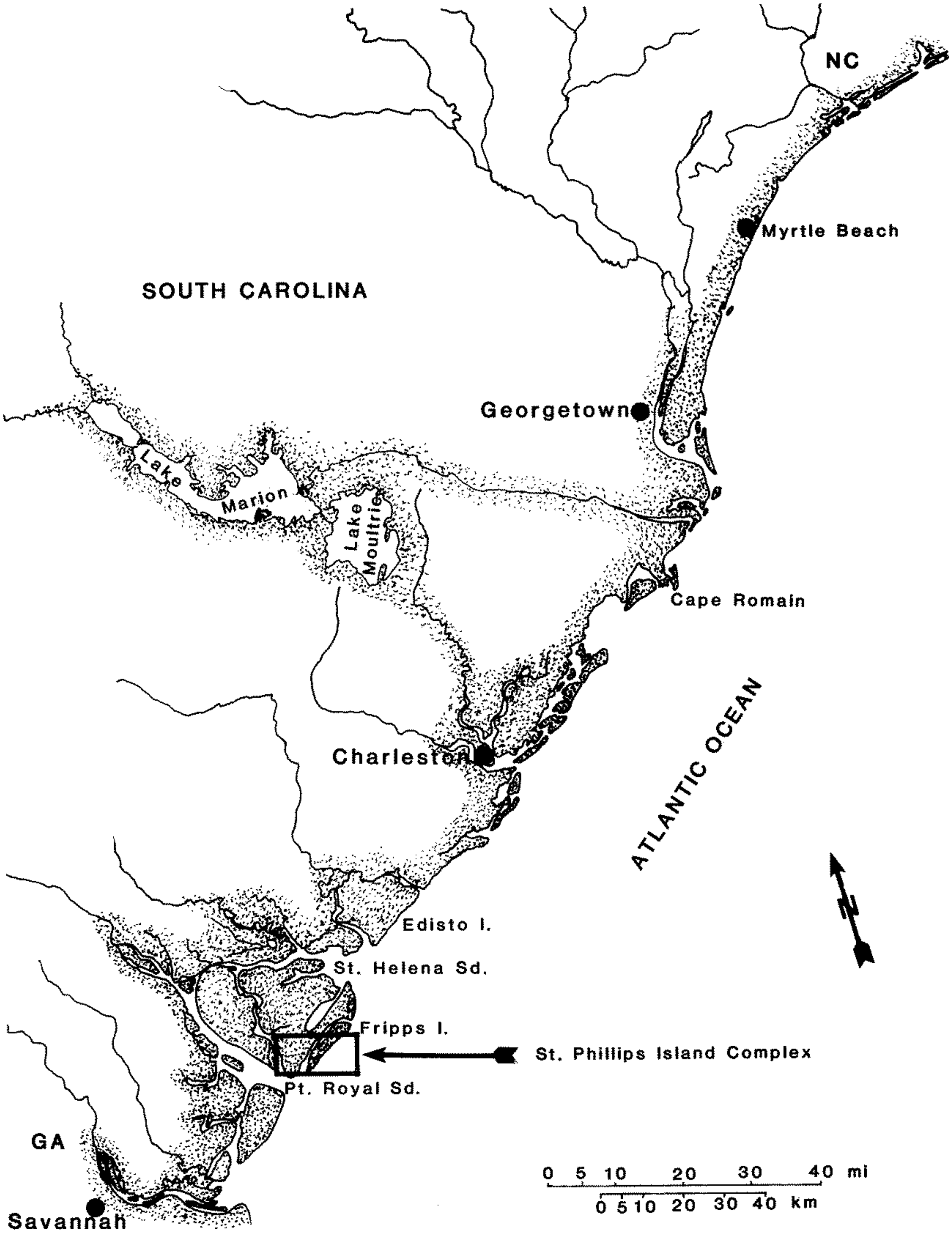

Figure 20. Location of St. Phillips Island Complex.

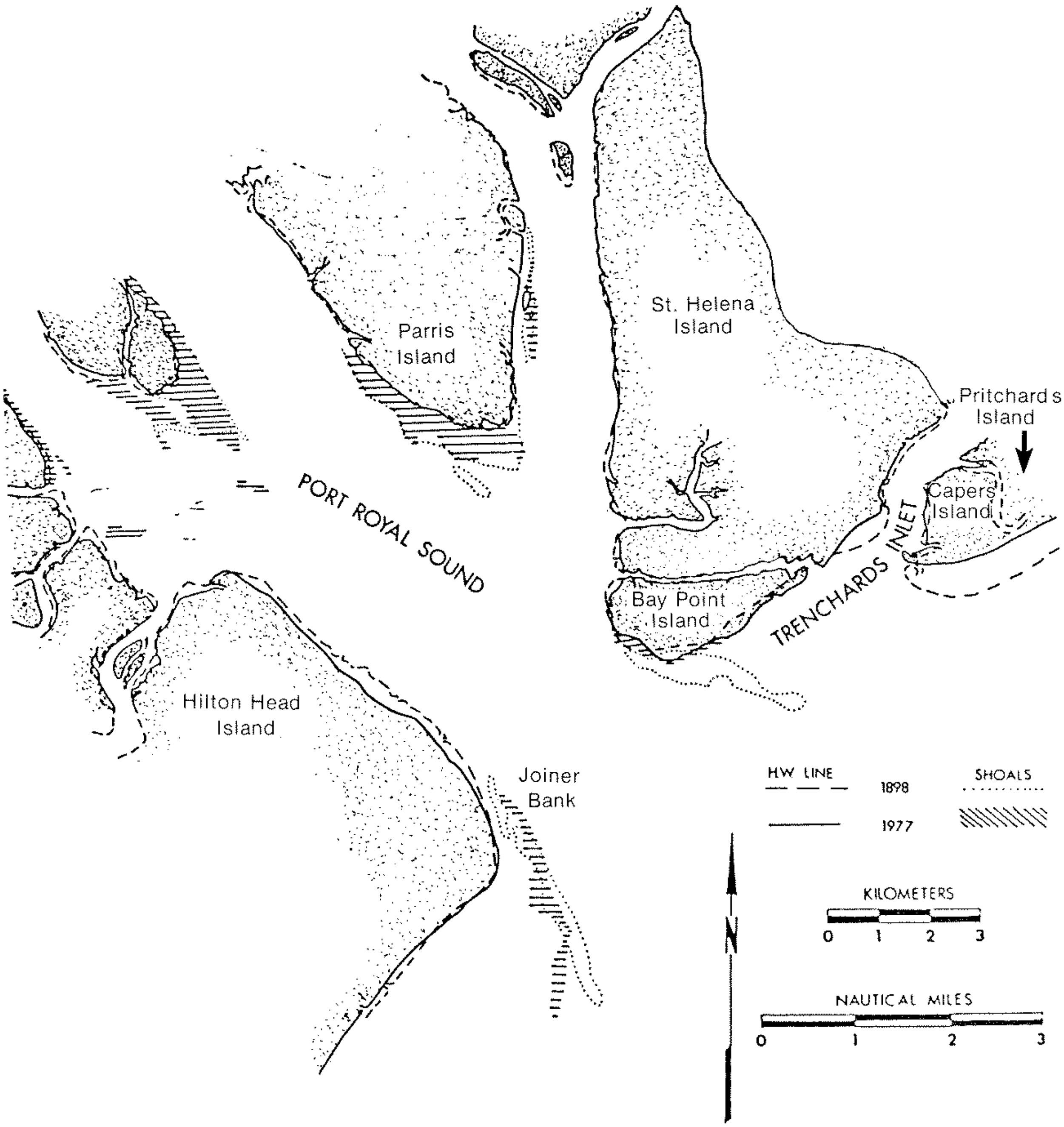

Figure 21. Recent shoreline changes for Port Royal Sound (Zarillo et al. 1984).

Undoubtedly, St. Phillips and Bay Point Islands have been strongly influenced by sedimentation processes at the mouths of Port Royal Sound and Trenchards Inlet, while Pritchards and Capers Islands have been influenced by adjacent tidal inlets producing alternating phases of shoreline transgression (erosion) and regression (progradation).

Modern Changes

Historical maps and charts have been analyzed to reconstruct the recent history of the St. Phillips Island Complex (Figure 21). This area has experienced dramatic shoreline changes (erosion and accretion) as is shown in Figure 22, and summarized below (Hayes et al. 1979a; Zarillo et al. 1984).

1. The shoreline along St. Phillips Island, including Bay Point Island, has been highly unstable since 1859. During this time there have been several short-term periods of erosion and accretion (Figure 22). The overall net trend, however, has been one of erosion along the northeastern portion of the St. Phillips Island shoreline and seaward accretion along the southwestern portion and along most of Bay Point Island (Figure 22). The erosion-deposition patterns observed are apparently a result of lateral migration of Trenchards Inlet to the west, and sediment bypassing at the inlet's ebb-tidal delta (Figure 21). The magnitude of net shoreline change along the St. Phillips-Bay Point Island shoreline has varied from accretion of 328 ft to erosion of 1,312 ft since 1859.

2. Shoreline instability at the southernmost tip of Bay Point Island seems related to sediment dispersal patterns and tidal current processes at the entrance to northeastern Port Royal Sound. Over 328 ft of shoreline along this part of Bay Point Island have been converted from an emergent barrier island to an intertidal to subtidal barrier shoal over an 80-year period (Figure 21).

3. From 1898 to 1977, most of Capers Island experienced extreme erosion. Net shoreline erosion along the southern and northern portions of the island exceeded 1,640 ft (Figure 22). The central portion of Capers experienced a period of significant accretion from 1859 to 1933 (Figure 22), but has rapidly and continually eroded since that time to the present at rates in excess of 95 ft/yr (Hayes et al. 1979a).

4. Pritchards Island has been experiencing almost continual erosion along the entire length of its shoreline (Figure 22, profiles F-6, F-8, F-10). Net erosion since 1859 is approximately 1,476 ft for most of the island.

Coastal Change Processes

The dominant agents of coastal change in the area of the St. Phillips Complex are (1) waves, (2) tides and tidal currents, (3) longshore sediment transport, and (4) sediment bypassing at ebb-tidal deltas. Overall trends of shoreline erosion and deposition in this area are influenced heavily by sedimentation patterns around the numerous tidal inlets, especially Trenchard's Inlet and Port Royal Sound.

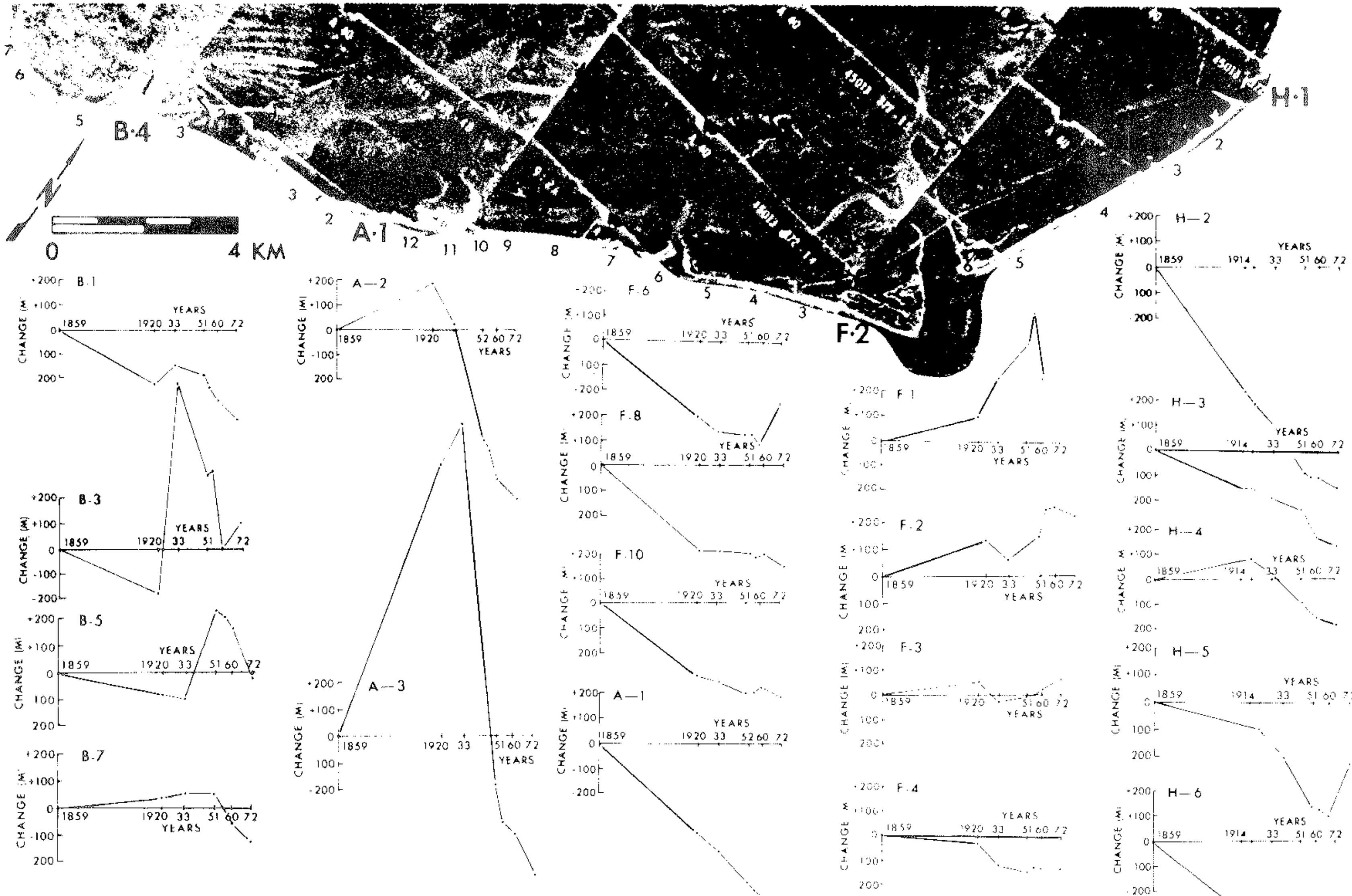

Figure 22. Erosion-deposition curves for sites along the South Carolina coast, including St. Phillips Complex. Sites are located as follows: F-6 through F-10 on Pritchards Island; F-11 through A-3 on Capers Island; B-1 and B-2 on St. Phillips Island; B-3 through B-7 on Bay Point Island (Hayes et al. 1979a).

Tidal range in the St. Phillips Complex area is near the maximum found in South Carolina. Mean tidal range is from 7.2 to 8.2 ft and spring tidal range may exceed 9.8 ft (Zarillo et al. 1984). Tidal current velocities can exceed 6.5 ft/s and are responsible for transporting large quantities of sediment in the areas adjacent to tidal inlets or at the mouth of Port Royal Sound. Large subaqueous to intertidal shoals, whose dimensions strongly affect the adjacent barrier island shoreline, are associated with the tidal inlets in this area, especially with Trenchard's Inlet and with Port Royal Sound. The relative size of these shoals in the St. Phillips Complex area can be seen in Figure 21. The shoals, or ebb tidal deltas, can serve as both sources and sinks of sediment, and also act to refract approaching waves, focusing their energy on various segments of the shoreline and thus causing extreme erosion.

Management Implications

The historical data clearly documents the highly unstable nature of the St. Phillips Island Complex shoreline. Because of the strong influence that Port Royal Sound and the numerous tidal inlets in the area have on coastal change, the adjacent shoreline has historically experienced dramatic short-term periods of erosion and deposition. Historical data strongly suggest that even those areas that are presently experiencing net accretion are probably in a short-term (spanning a few years) depositional phase that will be followed by erosion. While landward migration of these islands is not as dynamic a process as it is for many other barriers, shoreline instability and fluctuating erosion/deposition trends (over a period of 10 to 20 years) are quite common.

The St. Phillips Island Complex is under little development pressure because it is remotely located, it is low in elevation and subject to frequent flooding, and there are still ample topographically higher areas in the region outside of the CBRS unit available for development.

This CBRS unit has a high fish and wildlife value; the wetlands are especially valuable as nursery grounds for recreational and commercial fisheries. The unit is generally unaffected by development (except for some camps and other minor development) and shoreline control structures.

CBRS UNIT N01--LITTLE TYBEE ISLAND, GEORGIA

Geomorphology

Little Tybee Island forms the southern flank of a cuspate foreland, or deltaic barrier complex, of the Holocene Savannah River Delta (Figure 23; Griffin and Henry 1983). Little Tybee Island is bounded to the south by Wassaw Sound and separated from Tybee Island to the north by Tybee Creek. Tybee amd Little Tybee are the two northernmost barrier islands on the Georgia coast. The CBRS unit includes the entire area of Little Tybee Island.

In comparison to most barrier islands on the Atlantic coast, Little Tybee Island is relatively short (5.6 mi) and wide (4.0 mi maximum). The island contains a broad expanse of tidal flats and salt marsh, meandering tidal

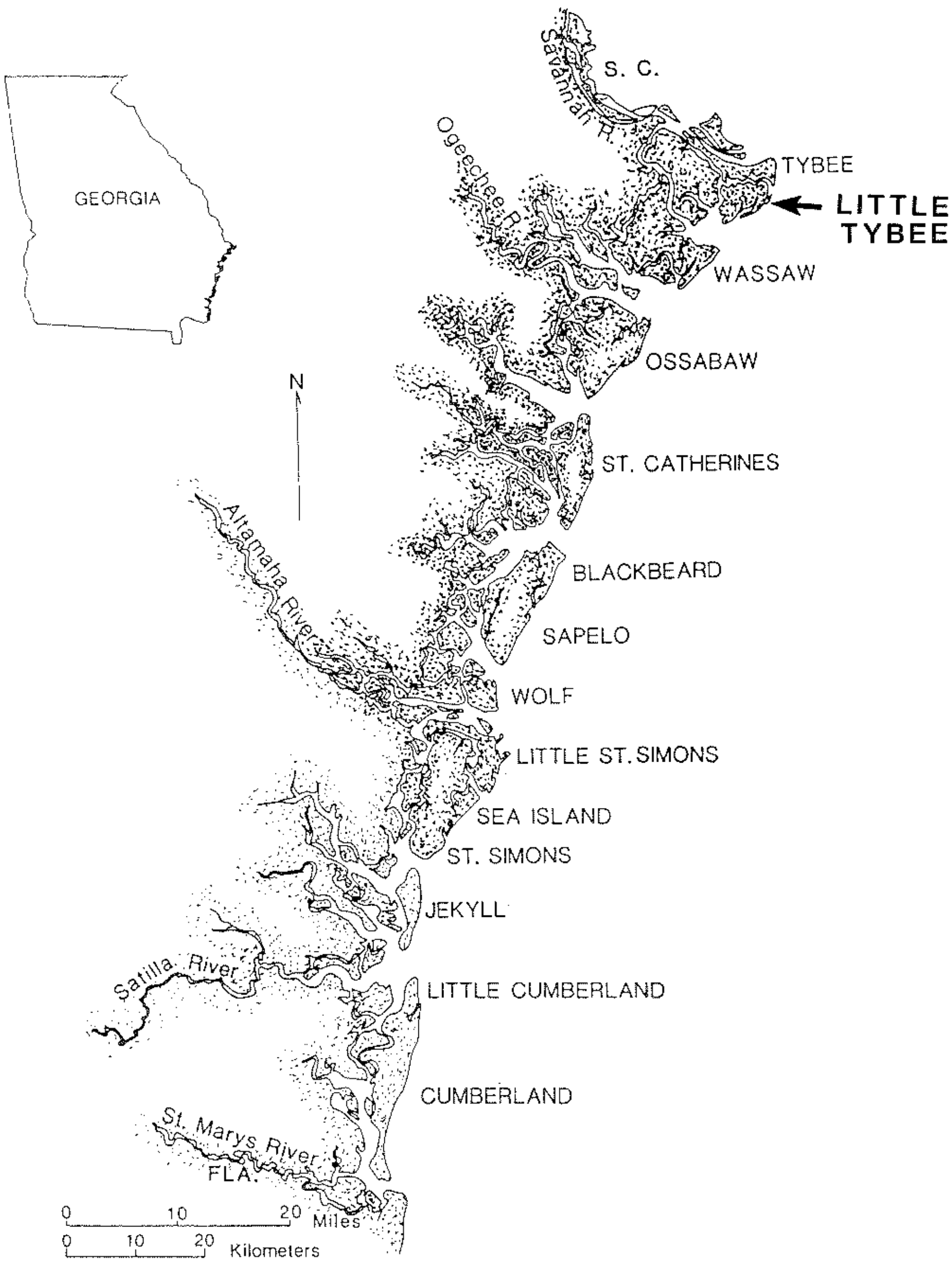

Figure 23. Location of the barrier islands of Georgia, including Little Tybee Island (Griffin and Henry 1983).

creeks, shore-parallel beach ridges, and flanking tidal inlets with large ebb-tidal delta shoals. The shape, morphology, and associated depositional environments of Little Tybee Island are characteristic of barrier islands in a mesotidal shoreline setting.

Little Tybee has a narrow, sandy beach, 5.0 mi in length (Mathews et al. 1980). Washover fans and terraces are common along the central and northern parts of the island and are indicators of the erosional processes that prevail there. An approximately 0.5 mi wide by 2.0 mi long belt of closely spaced shore-parallel to shore-oblique beach ridges occurs about 2.0 mi from the present beach. These beach ridges are presently surrounded by salt marsh, are interpreted to be Holocene in age, and apparently represent an earlier phase of progradation and deposition for Little Tybee. Elevations on the island range from sea level to 10 ft with nearly 90% of the land area covered by salt marsh (Warner and Strouss 1976). Most of the island is in a fairly natural state.

A 1.7 mi by 0.2 mi accreting sand spit known as Williamson Island occurs at the southern end of Little Tybee. Williamson Island formed between 1957 and 1960 when an inlet opened at the northern end of the spit to separate it from Little Tybee (Griffin and Henry 1983).

Geologic History

The origin and evolution of Little Tybee Island is directly related to the Holocene progradation and reworking of the Savannah River Delta. Little Tybee is separated from Wilmington Island, its Pleistocene counterpart, by a broad expanse of salt marsh and tidal flat deposits that overlie Holocene deltaic sediments associated with an earlier phase of progradation of the Savannah River (Hoyt and Henry 1971). Little Tybee and Tybee Islands were formed during a phase of transgression caused by rising sea level in the late Holocene. The islands are composed of sands derived from the reworking of older deltaic sediments. The islands are part of a cuspate foreland that probably represents an abandoned deltaic headland which began undergoing erosion and transgression approximately 3,000 years B.P. (Hoyt and Henry 1971).

Modern Changes

The recent history of all of Georgia's barrier islands was compiled by Griffin and Henry (1983) through an analysis of mean high water shoreline changes from 1857 to 1982. Their history of Little Tybee Island is based on available maps, charts, and air photos. It is presented below and in Figures 24-26.

Little Tybee Island has had a highly dynamic shoreline since 1866, with periodic advances and retreats occurring on the north, central, and southern portions of the island. The island underwent net accretion on all three segments during the period 1866-1913 (Figure 24). In the north segment, accretion on the northeast and southwest exceeded erosion to the southeast; there was a maximum advance of 2,300 ft on the downdrift end. A shifting pattern also took place on the central portion during this period, with a

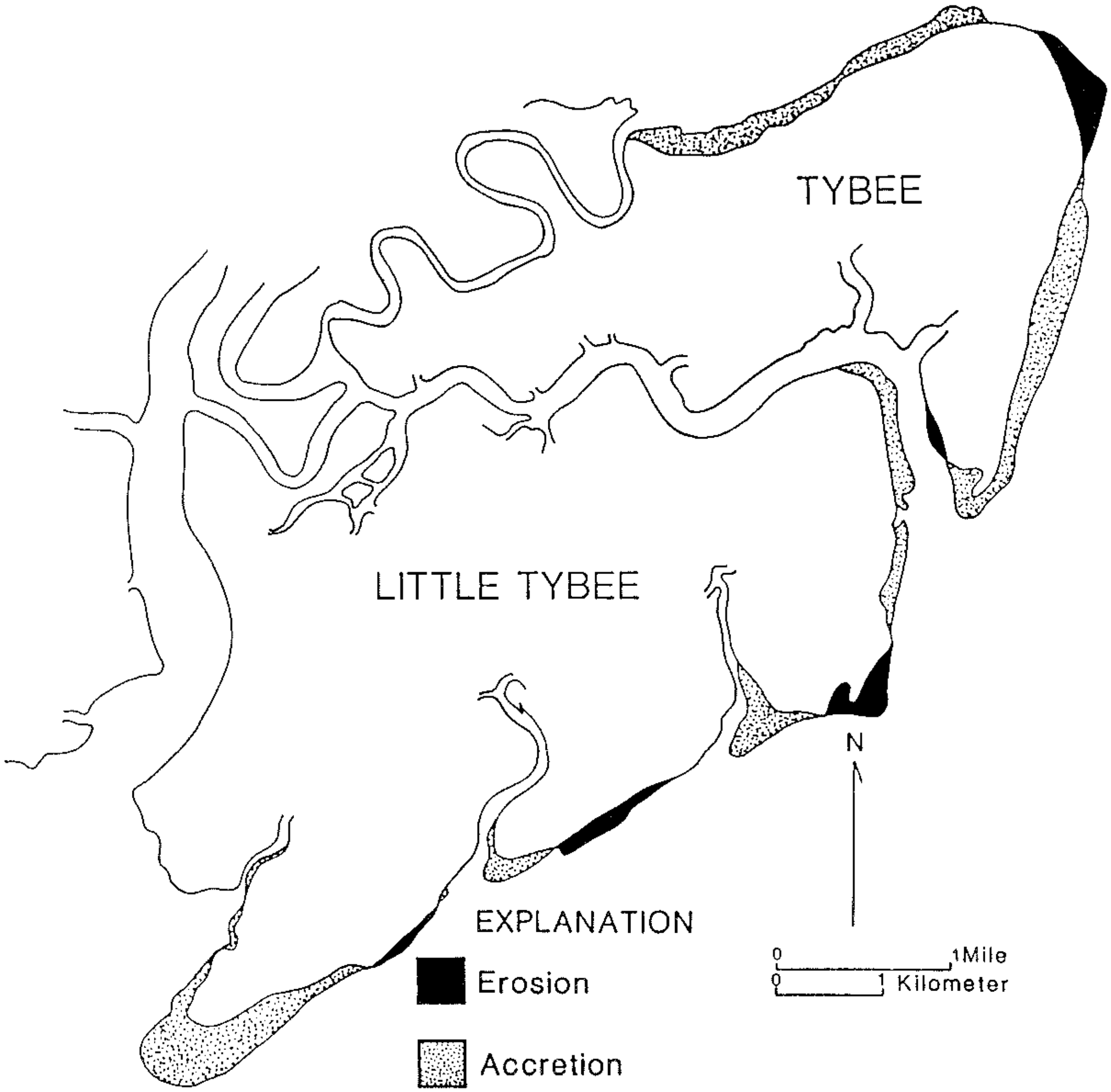

Figure 24. Net shoreline changes on Tybee and Little Tybee Islands, 1866-1913 (Griffin and Henry 1983).

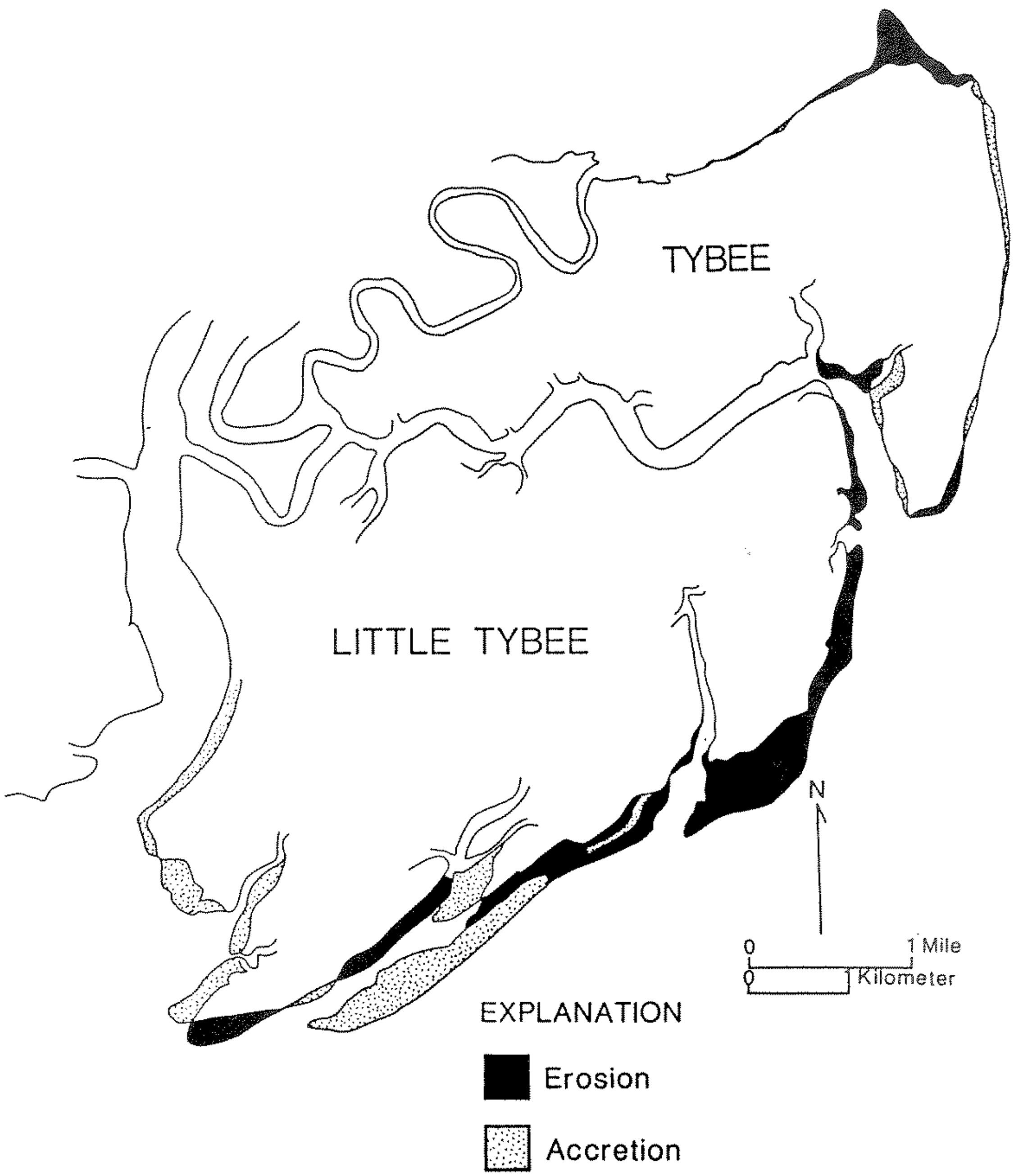

Figure 25. Net shoreline changes on Tybee and Little Tybee Islands, 1925-74 (Griffin and Henry 1983).

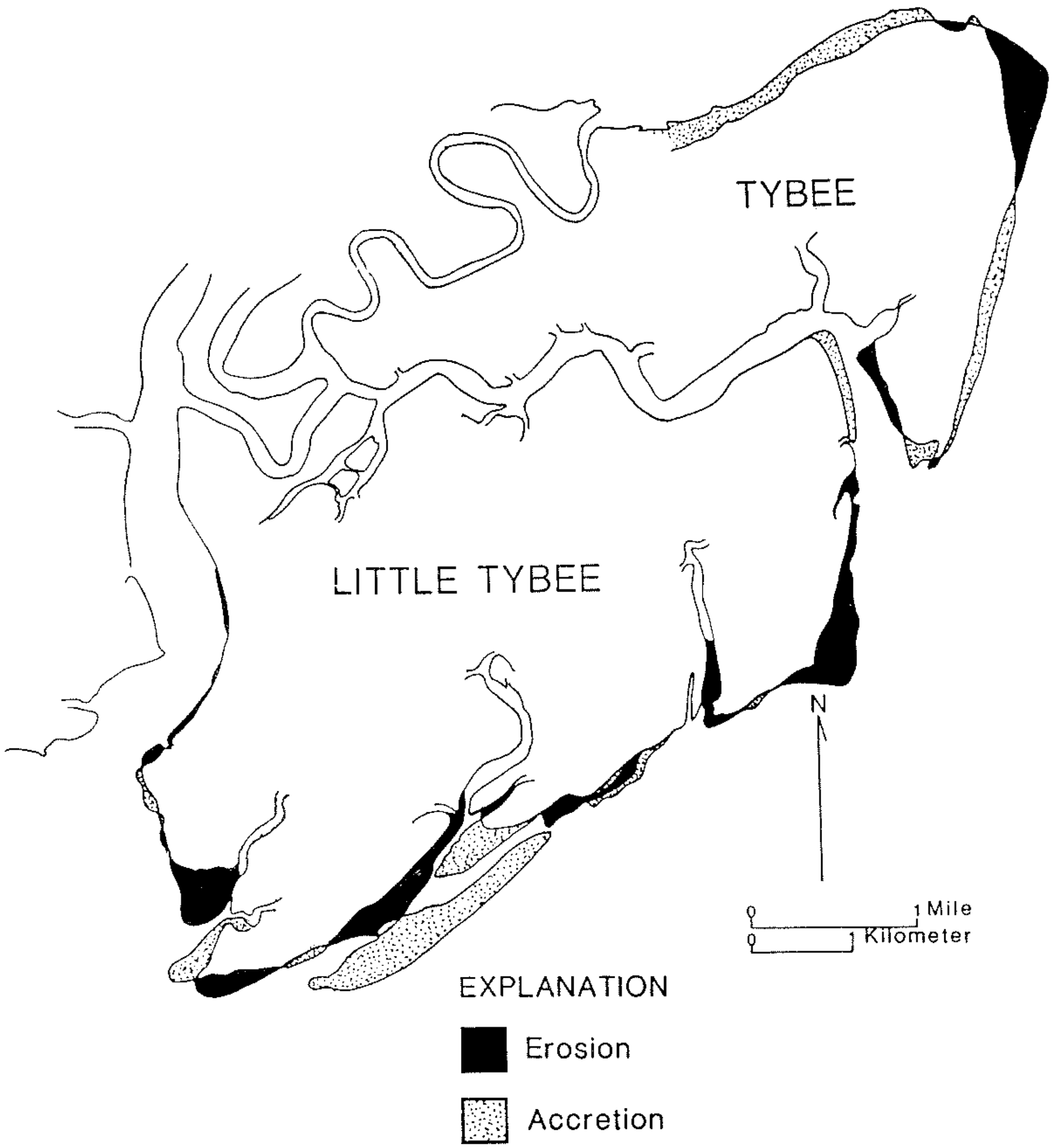

Figure 26. Net shoreline changes on Tybee and Little Tybee Islands, 1866-1974 (Griffin and Henry 1983).

maximum retreat of 440 ft and a maximum advance of 780 ft. On the Wassaw Sound shoreline, accretion took place at rates of 40 ft/yr.

Between 1913 and 1925, the north segment continued to accrete at rates of up to 100 ft/yr. The central section accreted all along the beachfront, with the average advance amounting to about 1,000 ft. On the south portion, erosion commenced at maximum rates of 183 ft/yr. In the interval between 1925 and 1957, the beachfront of the north section showed great instability, with a maximum advance of 1,000 ft and a maximum retreat of 1,400 ft. The southward migration of a spit attached to the north lobe progressed at an average rate of 172 ft/yr. Meanwhile, the central section eroded from 22 to 50 ft/yr, and the south segment eroded at rates of from 18 to 35 ft/yr.

Between 1957 and 1974, apparently prior to 1960, the elongated spit of the north segment was truncated and migrated south to become what is presently known as Williamson Island. Meanwhile, the central section showed a maximum advance of nearly 2,000 ft on the south end. The shoreline of the south segment was marked by stability and accretion, with a maximum advance of 380 ft.

When net change on Little Tybee Island for the period 1925-1974 is considered (Figure 25), it is apparent that the island has retreated; if the sandy beachfront alone is considered, losses are even greater. The greatest erosion for the period occurred on the north segment, where the maximum retreat was about 1,800 ft. Losses on the central section are balanced by gains on Williamson Island, an extension of the island's shoreline. On the south portion the sandy beachfront has eroded as much as 900 ft, although there were gains in the marsh to the southwest (Griffin and Henry 1983).

The mean high water line drawn from 1982 high altitude photos suggests stability and/or accretion for Little Tybee Island for the period of 1974-1982; this trend is probably the consequence of longshore transport of sediment from the renourishment project completed in 1976 on Tybee Island.

Coastal Change Processes

Processes of coastal change for the Georgia barrier islands include sea-level rise, tides, wind and wave climate, hurricanes and severe storms, and human impacts. As shown below, all of these processes have an effect, to varying degrees, on coastal change at Little Tybee Island.

Data from a tide gauge at Fort Pulaski, near the mouth of the Savannah River, indicate a rate of mean sea-level rise of 3 mm/yr during the 40-yr period 1936-1975 (Hicks 1978). This slow but fairly constant rate of sea-level rise is the greatest contributor to long-term coastal erosion and is the principal reason for barrier island migration.

Little Tybee Island is near the apex of a regional embayment along the southeast U.S. coast referred to as the Georgia Bight. The embayment amplifies tidal range so that the Little Tybee area experiences some of the highest tides along the entire U.S. east coast (Hubbard et al. 1979). Mean tidal range is 6.6 ft and spring tidal range averages about 10 ft (Griffin and Henry 1983).

The Georgia Bight is associated with a broad, shallow Continental Shelf that acts to reduce waves and wave energies approaching the shoreline. As a result, Little Tybee and most other Georgia barrier islands experience some of the lowest average wave heights (0.75-1.0 ft) along the U.S. east coast (Hubbard et al. 1979). However, significant coastal erosion due to wave processes has been recorded during extratropical storms approaching from the northeast and during hurricanes. Also, wave heights are locally amplified and focused by wave refraction around shoals or ebb-tidal deltas.

Hurricanes and northeasters are responsible for most beach and dune erosion and deposition of overwash on Little Tybee Island (Griffin and Henry 1983). Severe hurricanes have struck the Georgia coast on an average of once every 10 years. Extratropical storms are generally less intense, but much more frequent and of longer duration than hurricanes; therefore, they are a more formidable agent of coastal erosion on the Georgia barrier islands and along the U.S. Altantic coast in general.

Griffin and Henry (1983) have observed a reversal in the trend of accretion and seaward progradation from 1857 to 1925, and an increase in the rate of erosion since 1925 for Tybee and Little Tybee Islands. They attribute much of this erosional trend to dredging in the Savannah River and harbor. Dredging activity in these areas increased rapidly after 1915 and presently about 8 million yd^3 of sediment are dredged annually (Oertel 1977). Damming on the Savannah River in combination with dredging has resulted in dramatically decreased amounts of sediment reaching the nearshore zone and littoral drift system along Georgia's barrier islands. This sediment deficit is believed to be principally responsible for the recent erosion observed along Little Tybee Island.

Management Implications

Little Tybee should remain in its present natural state. Much of the island's surface area is covered by salt marsh that is not suitable for development and the environment here is extremely sensitive to development. Also, the beach and foreshore of Little Tybee is a very unstable and rapidly changing zone. This trend can be expected to continue in the future.

The construction of any groins or jetties along Tybee Island to the north will trap what little sediment remains in the longshore transport system, thus exacerbating the erosion problem at downdrift Little Tybee Island. If possible, sediment dredged from the Savannah River ship channel should be pumped to the south (downdrift) in the nearshore zone of Tybee and Little Tybee Islands. Beach nourishment along Tybee Island should continue because this indirectly supplies sediment to downdrift Little Tybee Island.

Although there is no development (as of 1982) on Little Tybee Island, some pressure exists to extend the resort development of Tybee Island southward. Tybee Island is extensively developed. There is also some pressure to develop open pit mining for phosphate deposits on Little Tybee Island (Griffin and Henry 1983).

Fish and wildlife habitat value is high and would change appreciably if resort development or mining of the island occurs. The large wetland area on the island serves as an important nursery ground for fisheries, many of which have commercial and recreational value.

CBRS UNIT P05A--MATANZAS RIVER, FLORIDA

Geomorphology

The Matanzas River Inlet region of northeast Florida is located 14 mi south of St. Augustine and 40 mi north of Daytona Beach. The Matanzas River extends from St. Augustine Inlet in the north to Ponce de Leon Inlet in the south via the Intracoastal Waterway (Figure 27). Matanzas Inlet is bordered on the north by Anastasia Island, on the south by Summer Haven, and on the west by Rattlesnake Island, the site of the Fort Matanzas National Monument. The CBRS unit is bordered by Summer Haven on the north and Marineland to the south. The entire CBRS unit area is less than 10 ft in elevation, and approximately half of this area is less than 5 ft above mean sea level.

Matanzas Inlet is the last unaltered inlet on the east coast of Florida. There is a substantial offshore bar in the mouth of the inlet that is transitory in nature, and appreciable shoals inside the inlet make it unsuitable for navigational purposes, except by small craft. Records indicate that the Spanish ships of the 16th and 17th centuries were able to navigate the inlet at high tide (Bruun 1966).

Four major habitats are present in the area surrounding Matanzas Inlet. Each habitat supports different types of vegetation, but all the plants are adapted to high temperatures, saline sands, strong winds, and salt spray. The locations of the four habitats in the Matanzas Inlet area are indicated on Figure 28. Each habitat is described below. Descriptions are taken largely from Burnson (1972) and Davis (1975).

The first of the four habitats occurs in older, more stable sections of the area. This habitat supports live oak, palmetto, and some southern red cedar. These older sections are found on the inner portions of Anastasia Island, in Summer Haven east of Highway A1A, and to some extent on the inner reaches of Rattlesnake Island.

The second habitat, consisting of palmetto scrub and some grasses, usually surrounds the first habitat. This habitat represents a transition region between the older, more stable areas and either salt marshes or beaches.

The third habitat, the salt marsh, comprises a large portion of the general area, occurring north and west of the National Park Service property on Anastasia Island and throughout most of the area west of the Intracoastal Waterway, except where dredge spoil has been placed along that waterway. The most abundant of the marsh grasses is *Spartina alterniflora*. This species

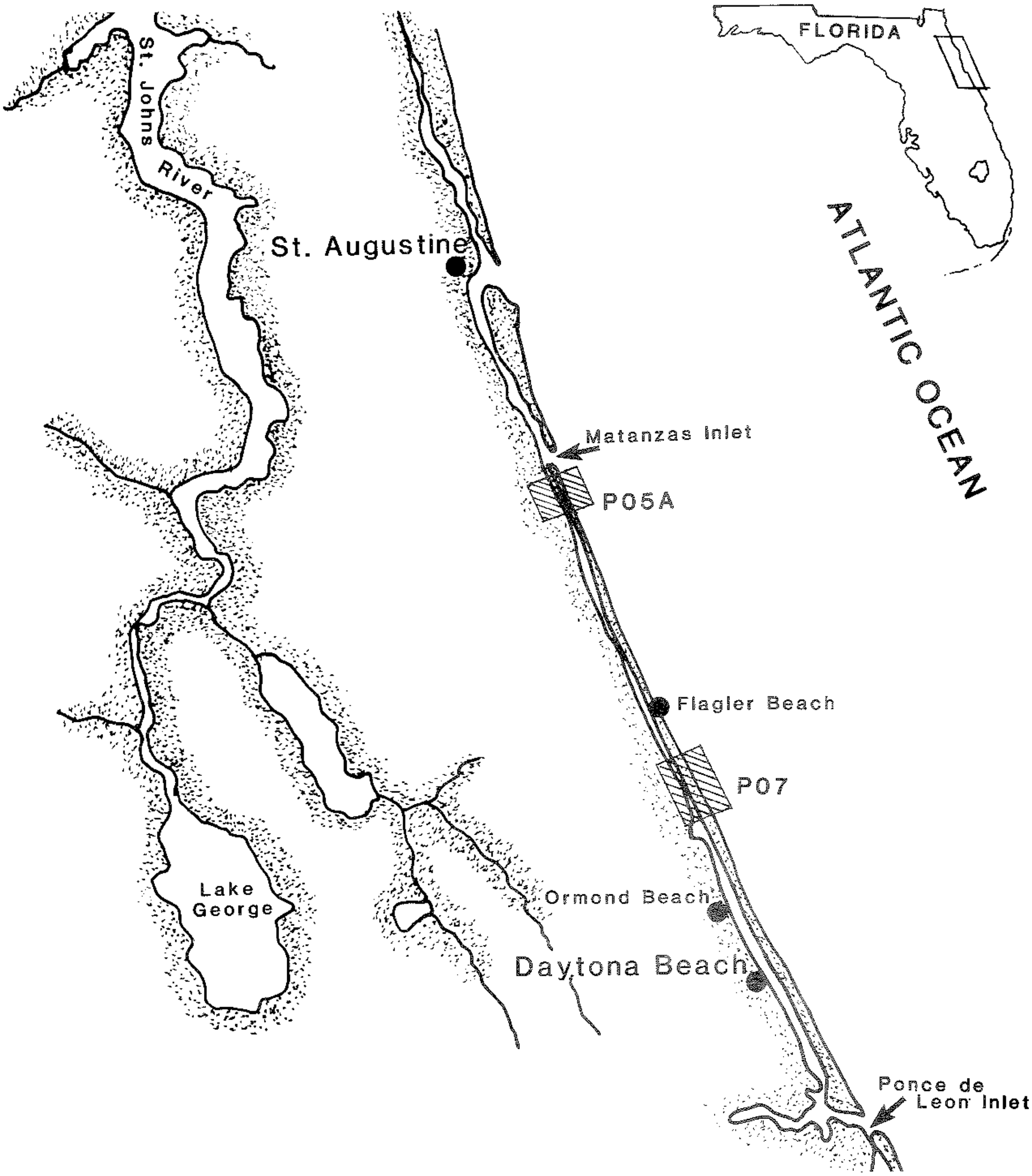

Figure 27. Location of CBRS units P05A, Matanzas River, and P07, Ormond-by-the-Sea, Florida.

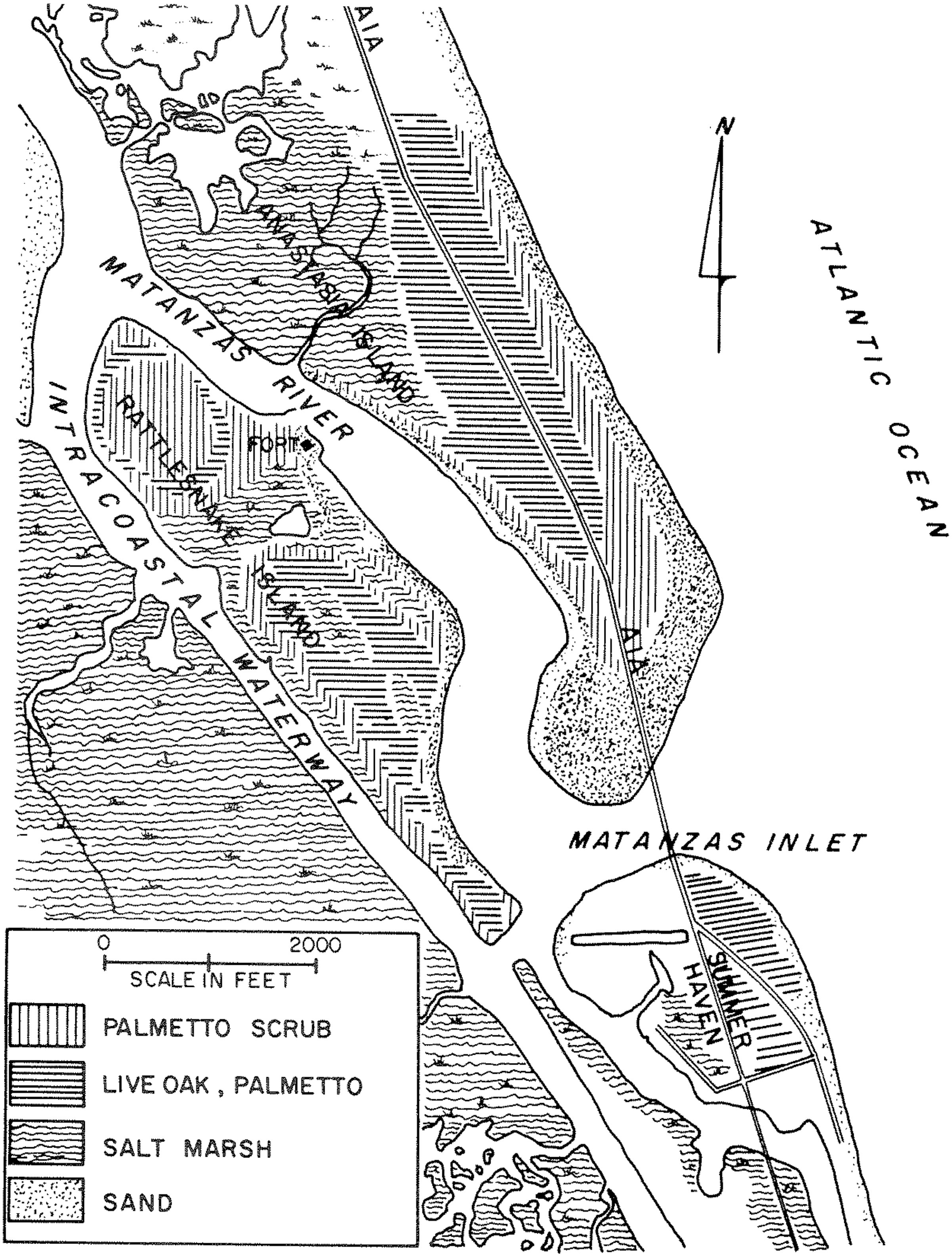

Figure 28. Vegetative base map of the Matanzas River area (U.S. National Park Service 1971).

stands from a few inches to a few feet in height. There are also some stands of the black mangrove (Avicennia nitidia) in the area, most occurring in the foremarsh along the Intracoastal Waterway. Some isolated mangroves grow to the west of the Intracoastal Waterway along the marshy sloughs and channels. Within the CBRS unit boundaries the wetland area is dominated by black mangrove.

The last habitat is the sand beach and dunes. The most noticeable plants in the inner sandy reaches are Yucca optunia and other small shrubs. On the dunes there are three predominant species: sea oats (Uniola paniculata), sea purslane (Sesuvium portulacastrum), and a small shrub called the marsh elder (Iva imbricata). All of these plants stabilize the dune areas. The roots of sea oats often extend downward as much as 10 ft, anchoring the dunes in place.

Geologic History

Underlying the unconsolidated sand along the east coast of Florida from Anastasia Island southward to the Palm Beach-Broward County line is a partially cemented rock formation known as the Anastasia. This formation is composed of different segments formed during several events in the Pleistocene (Brooks 1972). Outcrops of this formation appear along the Continental Shelf and are often found in locations where canals have been dug or inlets cut along the east coast of Florida. There are several exposed, consolidated outcrops on the beaches in the Matanzas Inlet area, as well as along various parts of the Intracoastal Waterway. One outcrop occurs on the southeast point of Matanzas Inlet and a larger outcrop, striking a northwesterly direction from the beach, occurs near Marineland.

The Anastasia formation varies from coarse rock composed of whole shells and minor amounts of quartz sand to a sandstone composed of carbonate and quartz sand particles. The cementing agent is either calcium carbonate or iron oxide (Cooke 1945). The surficial geologic structures in the Matanzas Inlet vicinity consist of perched barrier islands, which are Pleistocene features, overlain by mixed Holocene sands; a lagoon and tidal marsh area west of the barrier islands; and low elevation, low relief coastal terraces further to the west.

Core borings in the area indicate that the surficial sediments are composed primarily of a fine quartz sand with varying amounts of silt, clay, and shell mixed in (U.S. Army Corps of Engineers 1965). Offshore sedimentary characteristics were investigated between August 1966 and February 1967 by the U.S. Army Corps of Engineers. They used seismic reflection profiling and sediment cores to determine the availability of inner Continental Shelf sediments suitable for beach nourishment purposes. Results indicate that such material may be found offshore of Matanzas Inlet and Marineland. Detailed findings are available in the report by Meisburger and Field (1975).

Modern Changes

The major changes in the inlet and adjacent coastline from 1765 to 1972 are shown in Figures 29-31. It is apparent that the southern tip of Anastasia Island has migrated southward while erosion has taken place along the north

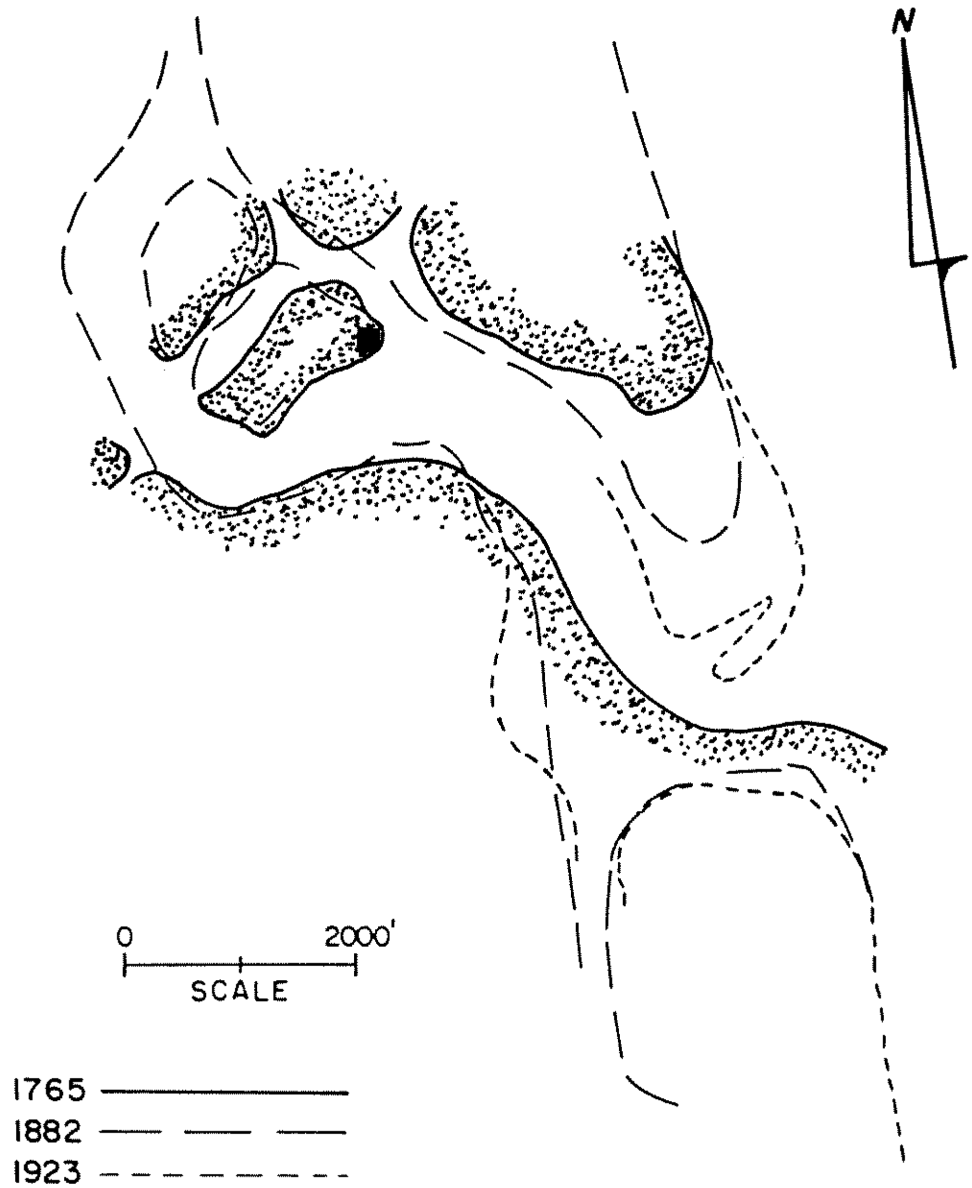

Figure 29. Shoreline changes around Matanzas Inlet, Florida, 1765-1923. Square indicates the location of Fort Matanzas (from Mehta and Jones 1977).

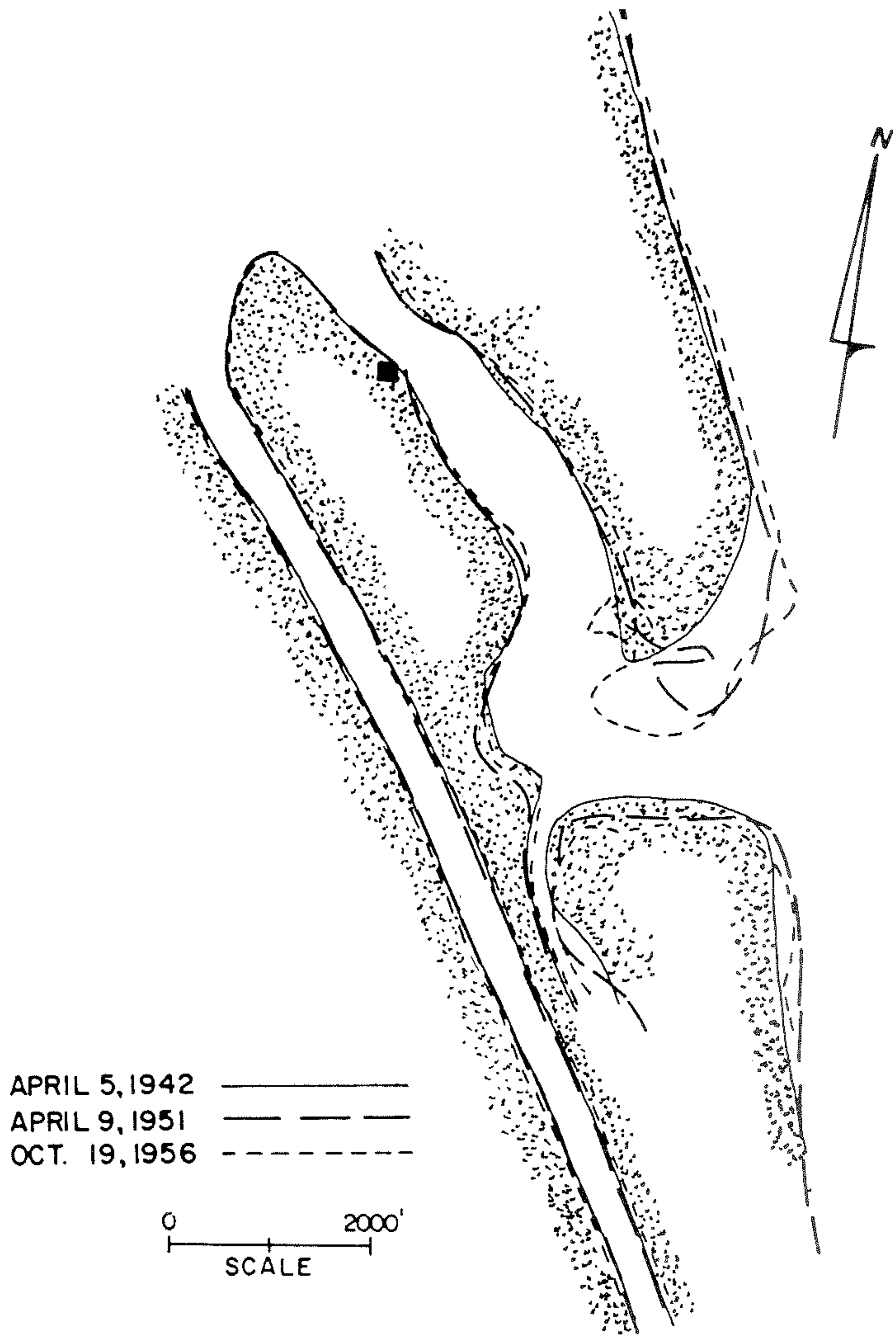

Figure 30. Shoreline changes around Matanzas Inlet, Florida, 1942-56. Square indicates the location of Fort Matanzas (from Mehta and Jones 1977).

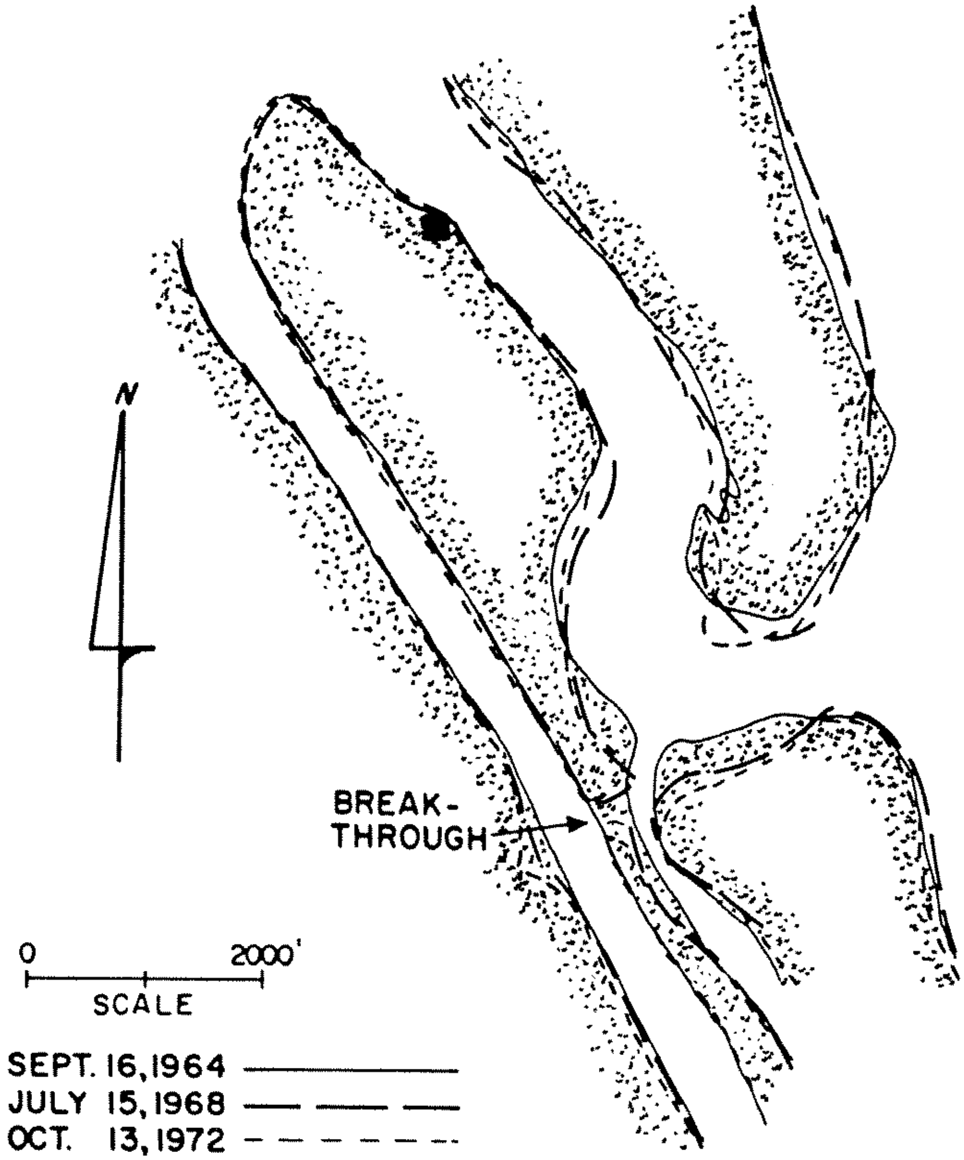

Figure 31. Shoreline changes around Matanzas Inlet, Florida, 1964-72. Square indicates the location of Fort Matanzas (from Mehta and Jones 1977).

and east sides of Summer Haven. The configuration of the present day Rattlesnake Island (Figure 29) has changed significantly, due almost entirely to the dredging in the Intracoastal Waterway. One interesting feature not shown on Figures 29 through 31 is Penon Inlet. It is said to have been just south of Summer Haven and to have closed in the early 1800's (Burnson 1972).

The Corps of Engineers (1965) calculated total and annual rates of accretion and erosion along the shoreline north and south of Matanzas Inlet. Over the period 1860-61 to 1963-64 the inlet widened. Erosion along the north shore averaged a moderate 1.4 ft/yr but along the south shore of the inlet erosion rates were approximately five times as great. Southward of the inlet from Summer Haven to Marineland (CBRS unit) erosion has been dominant but moderate, averaging approximately 1 ft/yr. Immediately north of the inlet (Anastasia Island) the shoreline has accreted an average of 0.8 ft/yr which is less than that for all of Anastasia Island. It is evident that the dynamics of Matanzas Inlet have affected the shoreline both north and south of it and that the general trend is one of erosion (see also Bruun 1962).

Coastal Change Processes

Along the St. Johns County coastline, northeast storms are, with few exceptions, more damaging than hurricanes. This is generally because the hurricane-generated winds and waves usually have a shorter duration and occur in a localized area, whereas a northeast storm may cause high winds and waves over a larger area for a longer duration. The typical northeast storm affecting St. Johns County is caused by a stationary high pressure area off the coast of the southeast United States with a low pressure area held directly south of the stationary high.

Storm tides of 2.5 ft above normal tides (mean tidal range = 5.2 ft) can be expected on an annual basis (U.S. Army Corps of Engineers 1972) and storm surges associated with major hurricanes can be expected to inundate the entire CBRS unit.

Between 1830 and 1968 there were 20 storms of hurricane intensity that passed within 50 mi of Matanzas Inlet--an average of 1 storm of hurricane intensity every 7 years. Between those same years, 46 storms of hurricane intensity passed within 150 mi of St. Augustine--an average of 1 every 3 years.

Apparent sea-level rise is a contributing factor to the coastal erosion problem. Apparent sea-level rise for this general area of the Florida Atlantic coast has averaged about 2.5 mm/yr during the past half-century (Hicks 1978).

Management Implications

This coastal region is beginning to experience development pressure from the north. The CBRS unit itself is bounded by two small communities--Summer Haven to the north and Marineland to the south. Further development of the area may lead to increased pressure for dredging and maintenance of Matanzas Inlet which will, of course, require continually increasing funding. Matanzas Inlet currently serves as a natural partial barrier to littoral sediment transport.

Dredging the inlet will probably increase its capacity to interrupt littoral transport, thereby further depriving the CBRS unit of a sediment supply and exacerbating erosion.

CBRS UNIT P07--ORMOND-BY-THE-SEA, FLORIDA

Geomorphology

Ormond-by-the-Sea lies on the east coast of Florida, about 9 mi north of Daytona Beach, in Volusia County (Figure 27). The CBRS unit is bounded on the south by the community of Ormond-by-the-Sea and extends nearly continuously northward to the Volusia-Flagler County line. The inland extent of this CBRS unit is the Intracoastal Waterway.

Beaches in this area average about 300 ft in width and lie in front of a low belt of vegetated dunes ranging in width from 650 to 2,600 ft (Shepard and Wanless 1971). The beach sand in the vicinity is clean, fine, and hard-packed, with a mean grain size of about 0.2 mm. The shell content is relatively low, but variable (Jones and Mehta 1978). Quartz in the beach sands was derived from South Carolina and Georgia, swept southward by littoral currents, while the shell content is derived from locally abundant outcrops of the Anastasia formation (Shepard and Wanless 1971).

The barrier beach complex is naturally separated from the mainland by the Halifax River, a saline tidal estuary. The Halifax River is connected to Mosquito Lagoon to the south, and opens to the Atlantic Ocean through Ponce de Leon Inlet (south of Daytona Beach). West of the Halifax River is an old beach ridge-swale complex on which about fifteen generally parallel dune ridges can be observed. This complex probably formed when sea level was relatively static during a Late Pleistocene interglacial stage. West of this belt is a wooded ridge representing a still older beach (Shepard and Wanless 1971). There are four major habitats in the area, all of which have been previously described for CBRS Unit P05A, Matanzas River.

Geologic History

The coast of Volusia County, like most of northeast Florida, is a low relief, low elevation coastal plain surface overlain by relict Pleistocene terraces and beach ridges (Meisburger and Field 1975). Outcrops of the Anastasia are easily weathered and eroded, providing shell fragments to the beaches along the coast. Pleistocene features on the barrier island beaches are usually draped with Holocene sands (Mims 1975).

Although Ponce de Leon Inlet is the only present-day inlet along the Volusia County coastline, geologic evidence suggests that some 1,500 years ago other inlets allowed direct drainage to the ocean. These inlets cut through the barrier island south of Ponce de Leon Inlet and east of Mosquito Lagoon (Jones and Mehta 1978). Within the boundaries of the CBRS unit no ancient or historic inlets are known, but detailed investigations have not been conducted.

Modern Changes

The Spanish first explored this area in the 1500's, initially naming the inlet Mosquito Inlet. They and subsequent Europeans found navigation to be very hazardous because of the rapid changes in channel positions within the inlet (Jones and Mehta 1978). An inlet stabilization project was completed by the U.S. Army Corps of Engineers in 1972 (Figure 32).

The jetties and weir constructed at the inlet have caused considerable change, and the morphology of the inlet is still adjusting to these structures (U.S. Army Corps of Engineers 1973; Jones and Mehta 1977). Although the inlet is considerably south of the CBRS unit the interruption of littoral transport of sediment by the jetties may be affecting the shoreline of the CBRS unit. The shoreline north of the inlet (including the CBRS unit) was erosional between 1936 and 1962 with rates generally less than 3 ft/yr (U.S. Army Corps of Engineers 1973). While detailed studies of shoreline changes in the vicinity of the inlet have been and continue to be conducted, largely under the auspices of the U.S. Army Corps of Engineers, the shoreline of the CBRS unit has received little attention.

Coastal Change Processes

Tides in the area range from 3.2 to 5.2 ft (Jones and Mehta 1978) but are not significant agents of coastal change in the Ormond-by-the-Sea area. Wave heights average less than 3.2 ft at an average period of about 8.5 seconds (U.S. Naval Weather Service Command 1970). Winds are predominantly out of the south and east during the summer months and out of the north during the winter months. The winds are typically 8 to 11 mph but are considerably higher during severe storms.

Hurricanes, tropical storms, and northeast storms cause flooding, beach erosion and related damages. Most of the damage is caused by flooding due to storm surge, wave runup and overtopping, and the undermining of structures as a result of erosion. Records show that a hurricane will pass within 50 mi of Ponce de Leon Inlet, on the average, once every 8 years, while northeast storms may occur several times during the winter.

Bruun (1962) postulated that the eustatic rise in sea level during the recent past has caused a general trend of erosion along Florida's coastline. Sea level trends over the past 50-60 years for several stations along the east coast of Florida show a rising sea level of 2.3 to 2.9 mm/yr (Hicks 1978). The eustatic component of this apparent sea-level rise has been estimated to be less than 2 mm/yr (Gornitz et al. 1982). In response to this rise, shoreline recession and beach steepening have been observed in varying magnitudes throughout the east coast of Florida. The coastline of Volusia County is eroding in response to this rise, while further south at Ponce de Leon Inlet, erosion is caused by the interruption of longshore drift by the inlet jetties.

Management Implications

This area is very near the famous resorts of Daytona Beach and New Smyrna and

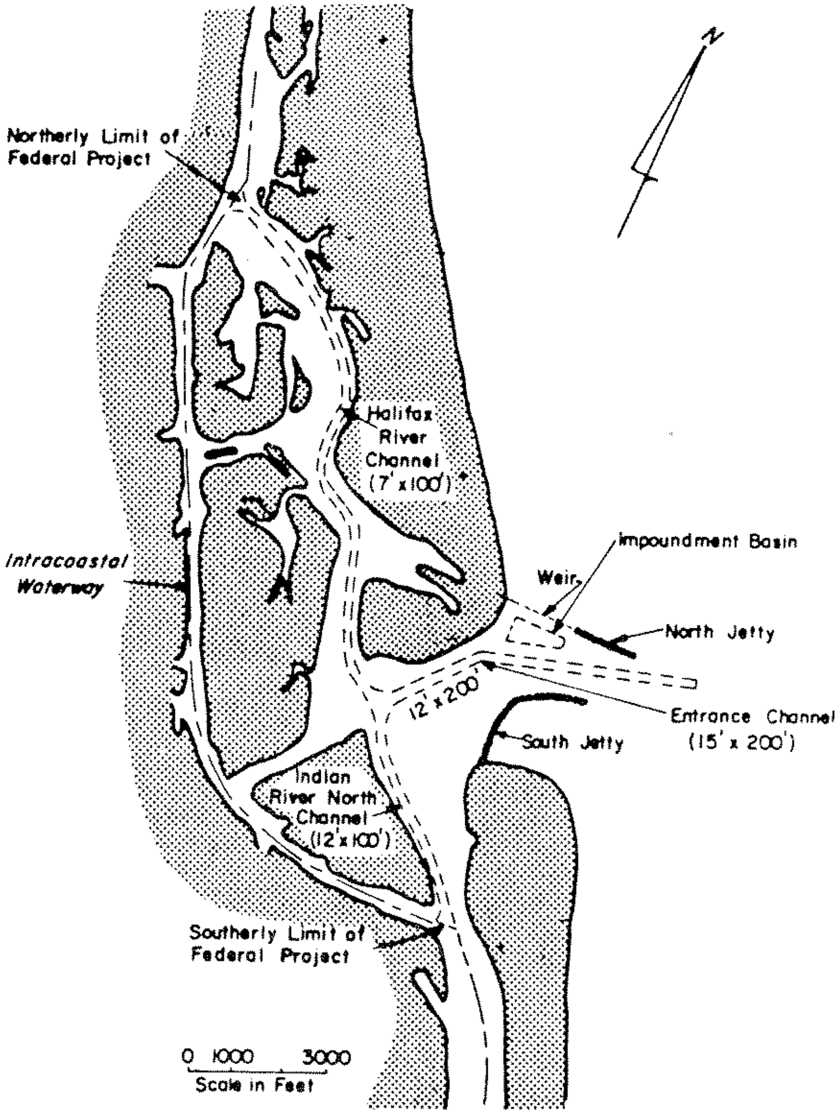

Figure 32. Federal inlet stabilization project at Ponce de Leon Inlet, Florida (Jones and Mehta 1978).

therefore, is under considerable development pressure. There is already considerable development immediately to the south of the unit, and an exit from Interstate 95 is only about 3 mi away. A monitoring program, including aerial photography and ground surveys, is needed to properly evaluate trends in shoreline behavior from Ponce de Leon Inlet to and including the CBRS unit. The interval between inventories should be short because of high development pressure and possible shoreline adjustments in response to jetty construction.

CBRS UNIT P11--HUTCHINSON ISLAND, FLORIDA

Geomorphology

Hutchinson Island stretches for 21 mi between St. Lucie and Fort Pierce Inlets (Figure 33). Most of the northern two-thirds (14 mi) of Hutchinson Island is included in the CBRS unit. The first 7 mi of the barrier island south of Fort Pierce Inlet are very low and susceptible to flooding. The elevation of the highway traversing the area is, in most places, less than 5 ft above mean sea level. The barrier beach in this section of Hutchinson Island is influenced strongly by the inlet and is very narrow except for the 1.3 mi of recently restored beach just south of the inlet (just north of the CBRS unit along a developed section of the island). There is no dune line as such in this area, although a reasonably heavy growth of sea strand vegetation thrives landward of normal wave action. From approximately 7 mi south of Fort Pierce Inlet, a dune line approximately 10-15 ft above mean sea level starts and runs southward, paralleling the coast to within 1.5 mi of St. Lucie Inlet. The higher portions of the dune line lie nearer to St. Lucie Inlet. The seaward face of the dune is steep and the beach is low. Southward to St. Lucie Inlet no dune line exists and maximum elevations are 5 to 10 ft above mean sea level. The shoreline within the CBRS unit is largely undeveloped, and the lagoon side of the island is dominated by mangroves.

Coquina rock (Anastasia formation) appears at several places as a submerged reef that generally parallels the shoreline at various distances offshore, from the highwater line to 2,500 ft seaward. The coquina reefs dissipate a portion of the ocean's energy before it reaches the beach, and thus help to retard the rate of shoreline erosion. The disintegration of the coquina also provides an important source of beach material for the area (Purpura 1972). These features are especially prevalent near the entrance to St. Lucie Inlet. A reef lying off the entrance to the inlet is exposed at extremely low tides. About 2 mi north of the inlet, extensive worm rock and coquina formations at the shoreline (in the vicinity of House of Refuge Museum) have prevented waves from breaking through the barrier island into the Indian River.

On the mainland, about 3,000 ft inland from Hutchinson Island, sandhills 25-35 ft in elevation run continuously from Fort Pierce to St. Lucie Inlet. These high sandhills are sand dunes built upon old beach ridges formed during the Pleistocene. The sandhills and the almost 90 degree bend of the St. Lucie River suggest an ancient shoreline with a predominant southward littoral drift and an ancient inlet in the vicinity of Sewall Point.

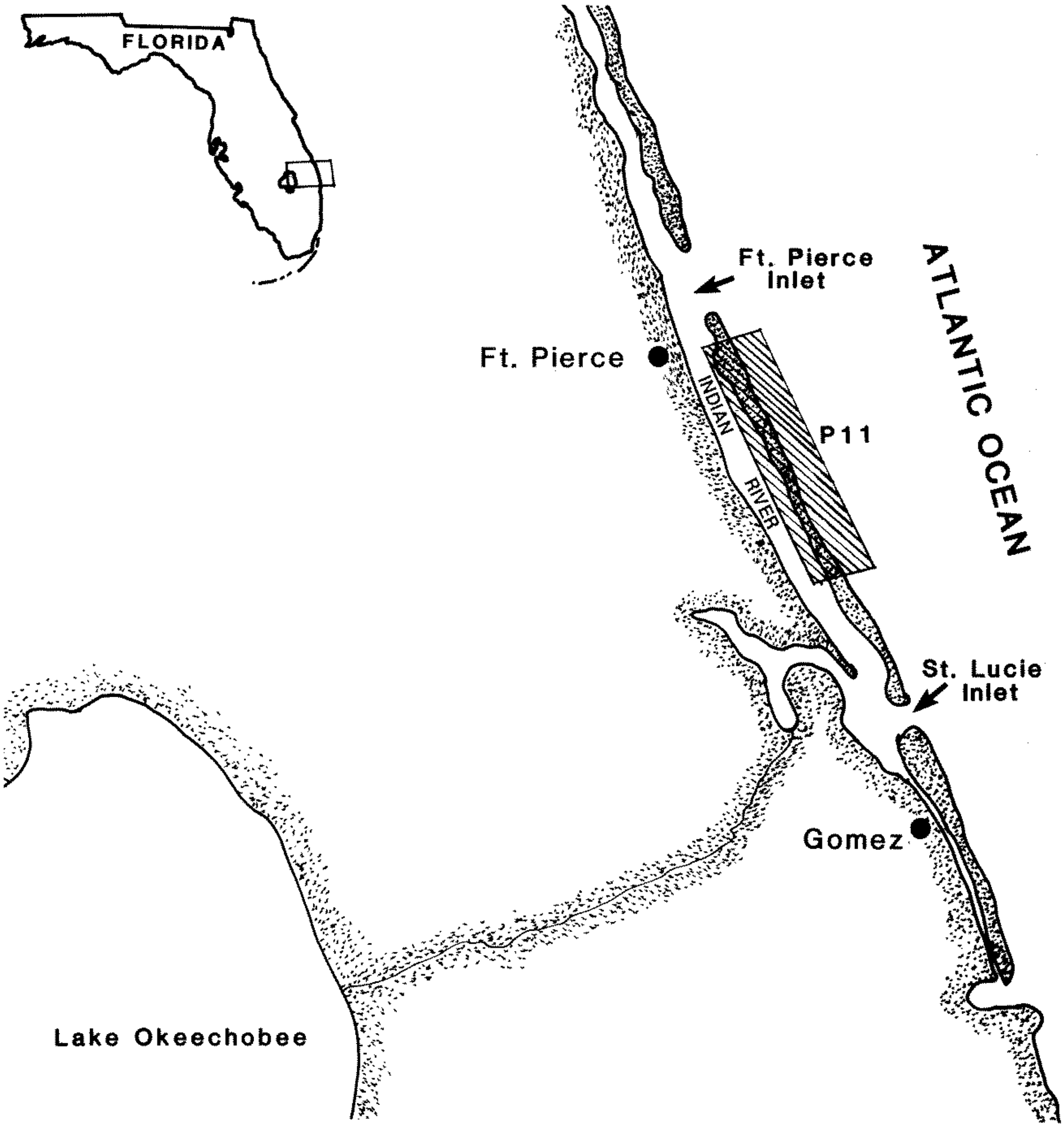

Figure 33. Location of CBRS unit P11, Hutchinson Island, Florida.

Geologic History

The barrier islands of Florida's east coast are thought to be part of an ancient offshore bar which emerged with upward tilting of the eastern part of the Florida Plateau (Shepard and Wanless 1971; Winker and Howard 1977). Rising sea level transported the barrier system landward. When relative sea level stabilized around 3,000 years B.P., wave and wind action became the dominant processes in modifying the barriers. The Pleistocene Anastasia formation underlies the entire CBRS unit and surrounding region (see Geologic History under CBRS unit P05A, Mantanzas River, for further discussion).

Modern Changes

The Fort Pierce area is unique in that erosion exists on the updrift side of the inlet where loss of sand (from many causes) outweighs accretion due to the north jetty (U.S. Army Corps of Engineers 1963b). Four major factors have contributed to a dominant erosional trend on the inlet's north side: (1) leakage of sand through the north jetty during periods of southward drift (the dominant direction of drift); (2) a longer north jetty than south jetty, which cuts off much of the sediment during periods of northward drift; (3) a gradual filling in of the old Indian River Inlet; and (4) apparent sea-level rise.

The shoreline for about a mile north of the inlet has generally advanced although outer portions of the profile have eroded, making the beach on the north side of the inlet generally steeper than would be expected, especially considering the amount of fine sand found in the profiles. The volumetric erosion rate of sand over the entire beach profile directly north of the inlet is 8,000 yd^3/yr/mi (U.S. Army Corps of Engineers 1971). During 1930-1957, the high water shoreline of the 1 mile section of shore directly north of the inlet prograded an average of 5.2 ft annually, but the nearshore bottom eroded to base rock, resulting in a steeper beach profile.

Erosion has also been a continuing problem on the south side of the inlet. The U.S. Army Corps of Engineers (1971) estimated a volumetric rate of erosion of 93,000 yd^3/yr for the 2.7 mi sector directly south of the inlet for 1930-1957. Unlike the north side of the inlet, the erosion occurred over the entire beach profile on the south side of the inlet. The average annual shoreline recession was 3-6 ft during this same period. The most severe erosion occurred approximately 1,200 ft south of the inlet where the shoreline has receded as much 450 ft.

Damage from northeasters and hurricanes was moderate in the area until 1962 (Bruun et al. 1962). A severe northeaster in March 1962 caused considerable erosion of the beach south of Fort Pierce Inlet. High breakers rolled over the section of the beach which lacked a dune line. Parts of the beach were reported to have been lowered by as much as 10 ft. The oceanfront road, which had an elevation 6 ft, had 0.5 ft of water over it during the height of the storm, and water entered homes along the road. The foundations of a few homes were undermined, and many homes had to be abandoned.

The St. Lucie Inlet area has had a history of both shoreline erosion and accretion north of the inlet (within the CBRS unit) and of continual erosion south of the inlet (beyond the CBRS unit). When the inlet was cut by natural processes in 1892, the shoreline experienced extreme erosion on both sides of the inlet. Historic surveys show that between 1882 and 1928 the shoreline for about 1.5 mi north of the inlet receded considerably, with a maximum recession of about 2,000 ft directly north of the inlet. The south side of the inlet experienced an even greater erosion problem. There are at least two reasons for the predominance of south-side erosion during the period of 1882 and 1928: (1) the southward-directed flows of the St. Lucie and Indian Rivers caused large southward nearshore velocities on ebb tides and (2) the inlet, functioning as a barrier to net littoral transport (recorded as being from north to south), trapped littoral drift material in a middle-ground shoal in a bar across the mouth of the inlet, and impounded material north of the jetty.

When the north jetty was constructed in 1929, the north shore of St. Lucie was stabilized, and accretion on the north side of the jetty took place. Between 1928 and 1946, accretion moved the shoreline seaward to a position in 1946 that approximately coincides with the 1882 position. In this same period, shoreline erosion continued on the south side of the inlet.

Between 1946 and 1962, the mean high water shoreline directly north of the inlet advanced 500 ft further, with an estimated annual accretion rate of 130,000 yd^3 in the 2.25-mi reach north of the jetty. The south shore continued to erode further during this 16-year period, with a maximum recession equal to 720 ft occurring at a point about 1.5 mi south of the inlet. About this time, the shoreline at the south point of the inlet reached an equilibrium position, while erosion continued further to the south.

Since 1962, the shoreline north of St. Lucie has continued to accrete while the shoreline south of the inlet has continued to erode. The continual erosion on the south side is indicated by the dead Australian pine trees which are present along the beach for a 3-mi stretch south of the inlet.

Hurricanes and northeast storms have caused considerable damage to the north on Hutchinson Island. Jensen Beach and Stuart Beach (just south of the CBRS unit) have experienced a considerable loss of sand during major storms. Storms in October 1963 and August 1965 (hurricanes), and in March 1962, December 1963, and January 1964 (northeasters) damaged or destroyed seawalls, retaining walls, and upland buildings and facilities. They also eroded the sand from the recreational beaches, removing sand from the beach to a depth of as much as 6 ft. The beaches and recreational areas were partially replenished when bulldozers, draglines, and trucks redistributed sand gained during favorable weather. Hurricane Betsy, in September 1965, completely eroded the sand from both Jensen Beach and Stuart Beach. However, natural recovery improved conditions considerably after the storm.

The beaches on the south portion of Hutchinson Island are probably not influenced by St. Lucie Inlet to the same extent as are the beaches on Jupiter Island further to the south because of the protective effect of the north jetty at St. Lucie. The erosional problems on south Hutchinson Island are probably primarily due to a net flux of sand being transported offshore during the natural onshore-offshore seasonal motion of sand, and to apparent sea-level rise (Bruun et al. 1963).

Coastal Change Processes

The tide near Hutchinson Island is semidiurnal with a large daily irregularity. The mean tidal range in the Atlantic Ocean is 2.6 ft and the spring tidal range is 3.0 ft. The tidal range in Faber Cove (northwest tip of Hutchinson Island) varies from 0.5 to 0.9 ft. The tide in the landward lagoon (Indian River) lags behind the ocean tide by about 2 hours, while maximum flood current in the inlets corresponds to high tide in the ocean.

Information on extreme tides in this area is sparse, but during the October 1953 hurricane, an ocean tide level of 6.3 ft was recorded by the U.S. Geological Survey at Eau Gallie to the north of Fort Pierce (Bruun et al. 1962; Bruun et al. 1963). Extreme tides measured in the Indian River (ocean tide unknown at corresponding times) occurred during the hurricane of September, 1938--7 ft above mean sea level at Melbourne--and the northeaster of March 1962--6.5 ft above mean sea level at Fort Pierce. These extreme tides have been estimated to be a 1-in-25-year (Bruun et al. 1963) to a 1-in-50-year (Harris 1982) event.

During 1900-62, a total of 17 hurricanes passed within a 50 mi radius of Fort Pierce. This is a hurricane frequency of 1 every 3.7 years. If the number of severe northeastern storms were added to the list, the total severe storm frequency would be considerably higher. Unfortunately, the effects of the hurricanes and extratropical storms on Fort Pierce Inlet and surrounding shoreline have not been well-documented. Hurricanes generally drive a great deal of sediment into inlets where it is trapped in shoals. They also can cause tremendous erosion along the shoreline due to the strong longshore currents and steep waves they generate. In the case of Fort Pierce Inlet, the jetties tend to restrict the flow of sand into the channel and inner recesses of the lagoon and consequently transfer the problem of sand loss to the downdrift side of the inlet (i.e., to the CBRS unit). The previously mentioned northeaster of March 1962 is the worst known storm with regard to erosion of Hutchinson Island beaches.

The inlet also provides an easy access route for flood waters and waves to reach the lagoon and Fort Pierce. In the storm of September 11-19, 1947, tides and waves entering the inlet overtopped seawalls normally 8 to 10 ft above the level of the Indian River, flooding streets along the waterfront. In the storm of August 24-29, 1949, many homes along the west shore of the Indian River were flooded.

Wind velocity records in West Palm Beach show that wind velocities are greater from the northeast sector than from the southeast sector, but that duration of wind and wind variability are greater from the southeast sector (Walton 1973). These wind patterns probably are also representative for the Fort Pierce-St. Lucie area. Offshore wind records are more important than local wind data because winds from offshore areas are primarily responsible for waves acting on the coastline. Offshore wind data recorded by ships off the mid-Florida coast indicate that the strongest winds are from the northern sector and the predominant winds are from the northern and eastern sectors, but that on the average, the percentage of time that winds blow from the northeast and southeast are about equal.

Littoral drift is strongly dependent on wave height and wave direction. When waves are from the north or northeast, littoral drift is southward. When waves are from the south and southeast, the direction is reversed. The predominant wave direction is from the northeast and net littoral drift in the Fort Pierce-St. Lucie area is southward. Walton (1973), using shipboard wave observation, has estimated total littoral drift as 334,000 yd^3/yr south and 281,000 yd^3/yr north; thus, a net drift of 53,000 yd^3/yr to the south. The predominant littoral drift from September through March is toward the offshore, and northward littoral drift predominates from April through August.

Management Implications

The area surrounding Hutchinson Island is primarily devoted to fruit and vegetable farming, cattle raising, recreational and commercial fishing, and the winter tourist trade. The majority of waterfront land surrounding St. Lucie and Fort Pierce Inlets and their tidal shores has been absorbed by fishing, tourist, and development interests.

Most of the northern and southern ends of Hutchinson Island beyond the CBRS unit already contain development. Despite the close proximity of developed areas and the presence of Highway A1A through the CBRS unit, the development pressure may remain relatively low because most of the CBRS unit is extremely low in elevation and covered with mangroves. Further, the presence of the Hutchinson Island Nuclear Power Plant near the mid-point of the CBRS unit (but not a part of the unit) may dissuade potential residents from building there.

The unit's extensive mangroves are valuable fish and wildlife habitat, both as nursery grounds for fish and as rookeries for herons and other water birds.

CBRS UNIT P30--CAPE SAN BLAS, FLORIDA

Geomorphology

Cape San Blas is the most prominent cuspate foreland in the Gulf of Mexico (Figure 34). Its maximum seaward extension is 50 mi south of the general east-west trend of the gulf coast in this vicinity. To the south-southwest of the cape, shoals can be traced seaward for an additional 15 mi. This association of extensive shoals with cuspate capes is common. The two arms leading to the cape are of unequal length, as is true also of the Hatteras (North Carolina) group of cuspate forelands. The shorter east arm measures 4 mi, whereas the north arm, St. Joseph Spit (the CBRS unit) is 17 mi long (Price 1958; Tanner 1964). The CBRS unit is continuous and comparatively large. It extends westward from Highway 98 on the mainland to Cape San Blas and then northward along St. Joseph Spit to the Florida State Park Boundary (about half-way along the spit).

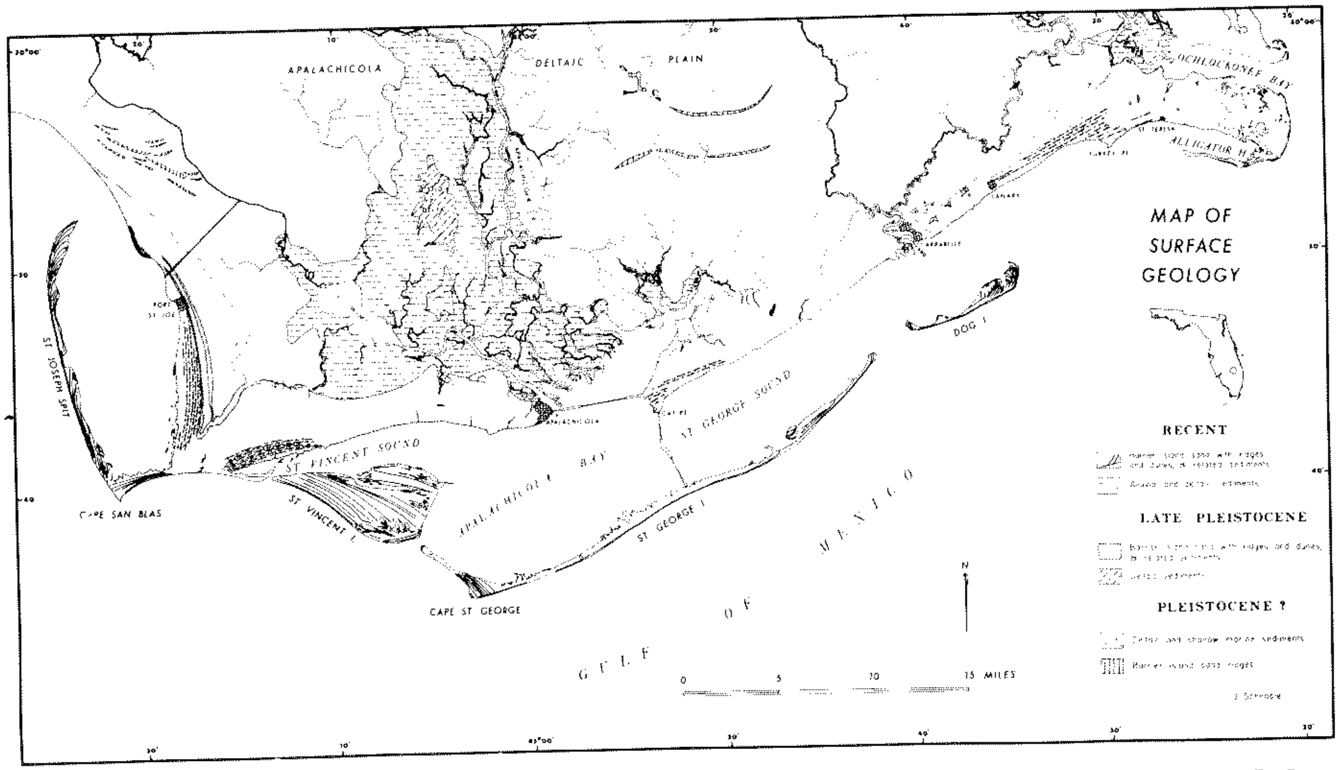

Figure 34. Surface geology of Cape San Blas and the Apalachicola River Delta (Schnable and Goodell 1968).

Marshy lowlands occupy much of the area between the Apalachicola River Delta, to the east, and the sandy shore areas of St. Joseph Bay and Cape San Blas (Kurz 1942). A few small sand ridges, probably old beaches, can be seen in the delta marshes northeast of Port St. Joe. These older sandy beach ridges seem to have grown southeastward, flanking the marsh and terminating about 6 mi north of the present shoreline. Successive ridges have overlapped the older ones, developing trends to the south and then to the southwest. The latest of the series of ridges from the eastern shore of St. Joseph Bay are truncated along the shoreline to the south.

Another series of ten ridges extends west-southwest, forming the north shore of Apalachicola Bay west of Indian Pass. The ridges are truncated by the present shoreline but may have joined the ridges southwest from Port St. Joe to form an earlier Cape San Blas, a mile south of the southeast corner of St. Joseph Bay. Following truncation of the old cape, sand drifted both east toward Indian Pass and St. Vincent Island and west to the present cape. This beach-drifting evidently established a new cape a little west of the present one, from which St. Joseph spit (the CBRS unit) began to grow northward. The beach ridges on the spit nearly all curve eastward toward St. Joseph Bay. As the terminal portion of St. Joseph Spit grew northward, the older, southern part suffered erosion as shown by the truncation of old ridges at the present gulf shore. At the north end, a series of overlapping hooks have developed.

Evidently, there is a rotary current in St. Joseph Bay, for sandy shoals in the southern part of the bay curve more sharply southwestward than does the bay shore (Curray 1965). Four breaks in these underwater sand ridges may be tidal channels. Thus, Cape San Blas, originating as arcuate ridges flanking the western and southern margins of the Apalachicola River Delta, has had a history remarkably well recorded by a series of beach ridges, chronicling growth interrupted from time to time by truncations.

Geologic History

Pleistocene sediments thicken from northeast to southwest along the coastal portions of the Apalachicola Delta region (Figure 34). Pleistocene sediments older than Sangamon age are not believed to exist. Two similar depositional cycles of Late Pleistocene sediments are best developed in the west and reflect sea level fluctuations prior to and during the Wisconsinan stage of glaciation. A series of radiocarbon dates from concentrated layers of wood that are believed to be associated with deltaic, shallow bay, and nearshore deposits indicate that this was an area of nearshore coastal sedimentation during parts of the late Pleistocene (Curray 1960, 1961). The old dunes and beach-dune ridges which exist on the mainland coast are considered to be Pleistocene bluff features (MacNeil 1949). Former shorelines and coastal configurations that existed during the late Pleistocene have had a strong influence on the modern shorelines and coastal configurations.

The Holocene rise of sea level is reflected in subsurface sediments of the former valley of the Apalachicola River, by the barrier islands, and by the offshore morphology of the Apalachicola Delta region. The rising sea first penetrated the river-cut valley approximately 9,950 years ago and first flooded it when sea level was about 75 ft below its present level (Shepard

1960). The Apalachicola Delta was located 5-10 mi north or northwest of its present position during the period of maximum flooding of the river valley, and the bay probably resembled many of the present bays lying farther to the west (Gorsline 1963). Sub-bottom profiles taken in Lake Wimico indicate that there are buried oyster bars under the existing lake bottom, which supports the hypothesis that much of the lower Apalachicola River valley was a bay area at one time (Kofoed and Gorsline 1963). The Apalachicola Delta has gradually prograded across the old bay floor as it has continued to build its way toward the gulf. The most noticeable progradation has occurred since present sea level was attained. However, a series of dams built along the Chattahoochee River during the 1950's and 60's has reduced the sediment supply to the Apalachicola Delta and will probably prevent the delta from prograding too much further into Apalachicola Bay.

Buried sediments in St. Joseph Bay indicate that no free circulation of gulf water existed at the time of deposition. This implies that there were barrier islands or bars which restricted the open exchange of gulf water. Although these bay sediments have not been dated, it is very likely that they were deposited at a slightly lower sea level than that which exists today. The morphology of the offshore shoal areas suggests that these features may have been barrier islands and that they were subsequently drowned with increasing sea-level rise.

Cape San Blas prograded over former nearshore sediments during the last part of the Holocene sea-level rise. A major distributary of the Apalachicola River then emptied into the Gulf of Mexico at a point approximately 5 mi north of the present St. Joseph's Bay (Stewart and Gorsline 1962). The offshore shoals to the west and south deflected the riverine and gulf waters, setting up gyral currents and causing westward spit growth (Cape San Blas) from the mainland. The spit grew and linked up with northwest-southeast trending offshore bars. Sediments were deflected by these bars which were transverse to the spit, resulting in deposition and the emergence of St. Joseph Spit (Stewart and Gorsline 1962; Schnable 1966).

Core borings on the cuspate foreland indicate that these Holocene sediments are almost entirely of barrier island origin. They consist of clean, fine- to medium-grained quartz sand with intermixed shell throughout much of each core. Most of the shell material consists of worn and broken fragments similar to those found in the swash zone and nearshore area of the islands today.

Cape San Blas is a classic example of a beach or spit deposit prograding over former nearshore deposits. The interval from 2.5 to 8.6 ft below the surface represents dune and beach deposits. These rest on obvious nearshore marine deposits with sediments and fauna similar to those existing today just west of the present cape.

Modern Changes

Historical shoreline changes in the Cape San Blas CBRS unit have not been the subject of any known investigations. GIS analyses presented in a later section of this report show that this CBRS unit has experienced an overall net gain in land area since the 1940's (Table 17). All of the land gain, however, is attributed to a relatively small portion of the CBRS unit. The shoreline at Cape San Blas (the physical cape, not the entire CBRS unit) has prograded seaward and the lighthouse which was close to the shoreline in the 1940's is now a considerable distance inland. The degree of shoreline progradation decreases eastward to the CBRS unit boundary where the shoreline has been fairly stable since the 1940's. From Cape San Blas northward along St. Joseph Spit, the shoreline is erosional, with the highest rates of erosion occurring at the southernmost end. At the northern end of the CBRS unit (about half of the total distance of the spit), the shoreline has been relatively stable to slightly accretionary since the 1940's.

The shoreline along the St. Joseph Bay side of the CBRS unit has been relatively stable except for some minor erosion along the northward-facing shoreline. This erosion is probably associated with waves generated by northerly winds following a frontal passage and the long fetch of St. Joseph Bay.

Coastal Change Processes

The CBRS unit is part of the much larger lower Apalachicola Basin. Sands in the region have been provided by the Apalachicola River. The reworking and redistribution of Holocene and Pleistocene quartz and shell sands through wave action and littoral drift provide the sand source for the CBRS unit (Stewart and Gorsline 1962). The large sandy shoal offshore of Cape San Blas is not regarded as an active source area for the present shoreline because the mean grain size of the shoal sands is too large for the low energy waves along this coast to transport this sand landward (Tanner 1964). The shoals themselves appear to have remained relatively unchanged for over a century (Schnable 1966).

Tides are mixed. Mean diurnal tidal range is 1.2 ft, and mean semidiurnal tidal range is 2.4 ft. Wave energies are low to moderate and littoral transport is approximately 200,000 yd^3/yr at Cape San Blas, decreasing to less than 100,000 yd^3/yr at the northern terminus of St. Joseph Island (Walton 1976).

This region has been tectonically stable since at least the late Pleistocene; thus, subsidence is not a contributing factor to erosional processes (Schnable 1966). Tide gauge data for stations to the east and west depict a rise in sea level approximately 1.4 mm/yr over the past several decades, which approximates estimates of eustatic sea-level rise (Gornitz et al. 1982).

Management Implications

The Cape San Blas CBRS unit was selected as a case history despite the lack of previous investigations examining shoreline change because it is currently under considerable development pressure. The general lack of shoreline change data for this area is uncommon compared to the other case studies presented here, but many of the CBRS units lack complete site-specific data sets.

Map and photo comparisons conducted for this study indicate that most of the gulf shoreline of this unit is eroding at fairly substantial rates, although the net change for the total CBRS unit area has been one of accretion over the past 40 years because of spit progradation at the cape. The high erosion rates along the shore make beachfront development a risky proposition.

The upper half of St. Joseph Spit lies outside of the CBRS unit, but is part of the State of Florida's park system. Florida has plans to acquire additional property south of the park (part of which is in the CBRS unit) as part of the "Save Our Coast" program.

The Cape San Blas unit has a high fish and wildlife habitat value. Older wooded beach ridges with intervening wetland swales provide a high degree of habitat diversity. On the St. Joseph Bay side of the CBRS unit, especially on the southern end of the Bay, fairly extensive seagrass beds are present. A small segment of the area bordering the CBRS unit is part of the St. Vincent National Wildlife Refuge.

CBRS UNIT Q01--MOBILE POINT, ALABAMA

Geomorphology

Mobile Point is the westernmost tip of the Morgan Peninsula, a large baymouth bar extending westward from the eastern shore of Mobile Bay (Figure 35). This peninsula, which varies from 0.17 mi to 2.2 mi wide and is 18 mi long, separates the bay and gulf waters, and ensures the maintenance of an estuarine environment in Mobile Bay (Hardin et al. 1976). The rectangular shape of Mobile Bay differs sharply from the dendritic pattern that characterizes most estuaries on the east coast of the United States. Mobile Bay was formerly twice its present length, but the upper part has been filled by the confluent Tombigbee and Alabama Rivers, which together drain much of Alabama.

The Morgan Peninsula spit is part of a Holocene ridge complex that stretches along the Alabama-Florida mainland shore (Shepard and Wanless 1971). The thickness and composition of different parts of this sequence were influenced by the topography and characteristics of the late Pleistocene land surface (Otvos 1985). Throughout the Morgan Peninsula, the thickness of the Holocene sediments reaches 46-65 ft, with individual ridge heights around 10-13 ft. More than 50 ridges occur, with multiple orientations (Figure 36). Recently some south-southwest trending ridges have been truncated by shoreline erosion.

A long underwater spit, known locally as Dixie Bar, extends about 3.7 mi south into the gulf from Mobile Point (Figure 37). Between 1929 and 1973, this spit has narrowed and become more elongate, and the southern tip appears to have moved slightly west. This indicates little or no deposition on the spit, either because of erosion by longshore currents or because very little material is being eroded from the gulf shore of the peninsula to the east (Hardin et al. 1976).

The CBRS unit makes up only a small part of the Morgan Peninsula. The CBRS unit is highly discontinuous. It consists of eight nonbordering subunits, five of which have no gulf shoreline and three of which encompass only a very small percentage of the gulf shoreline of the peninsula.

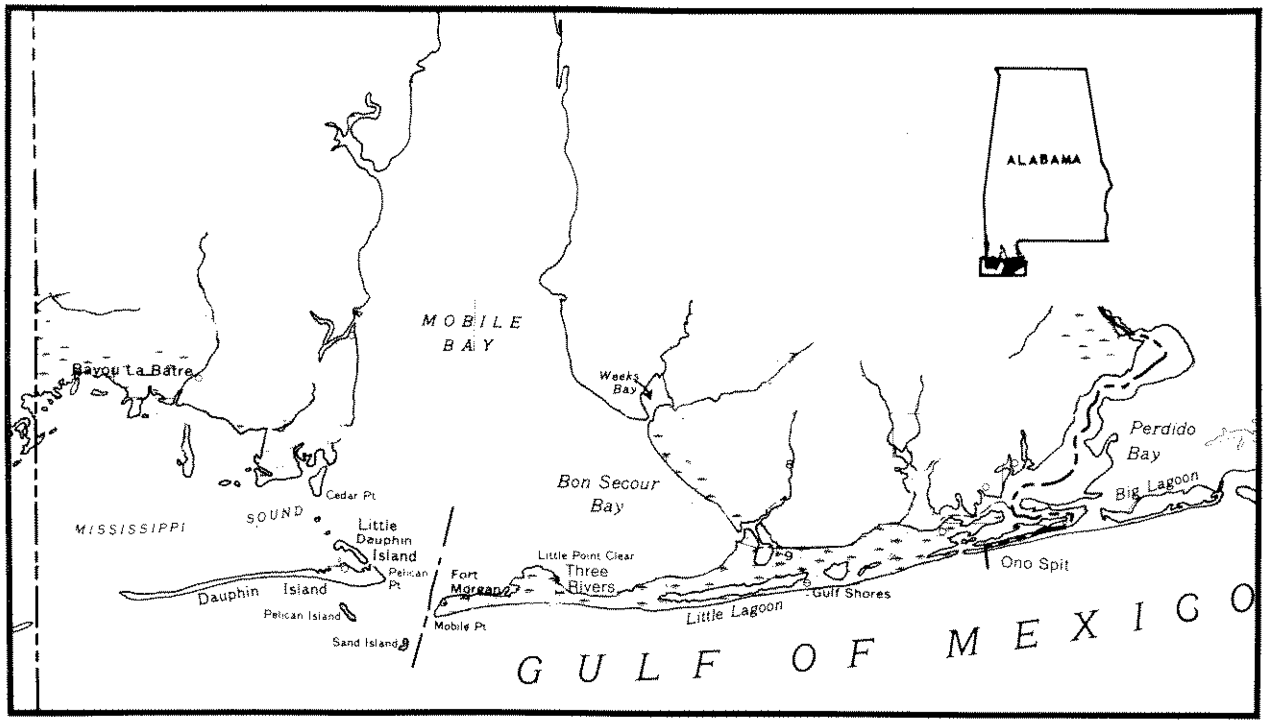

Figure 35. Location of Mobile Point and Dauphin Island, Alabama.

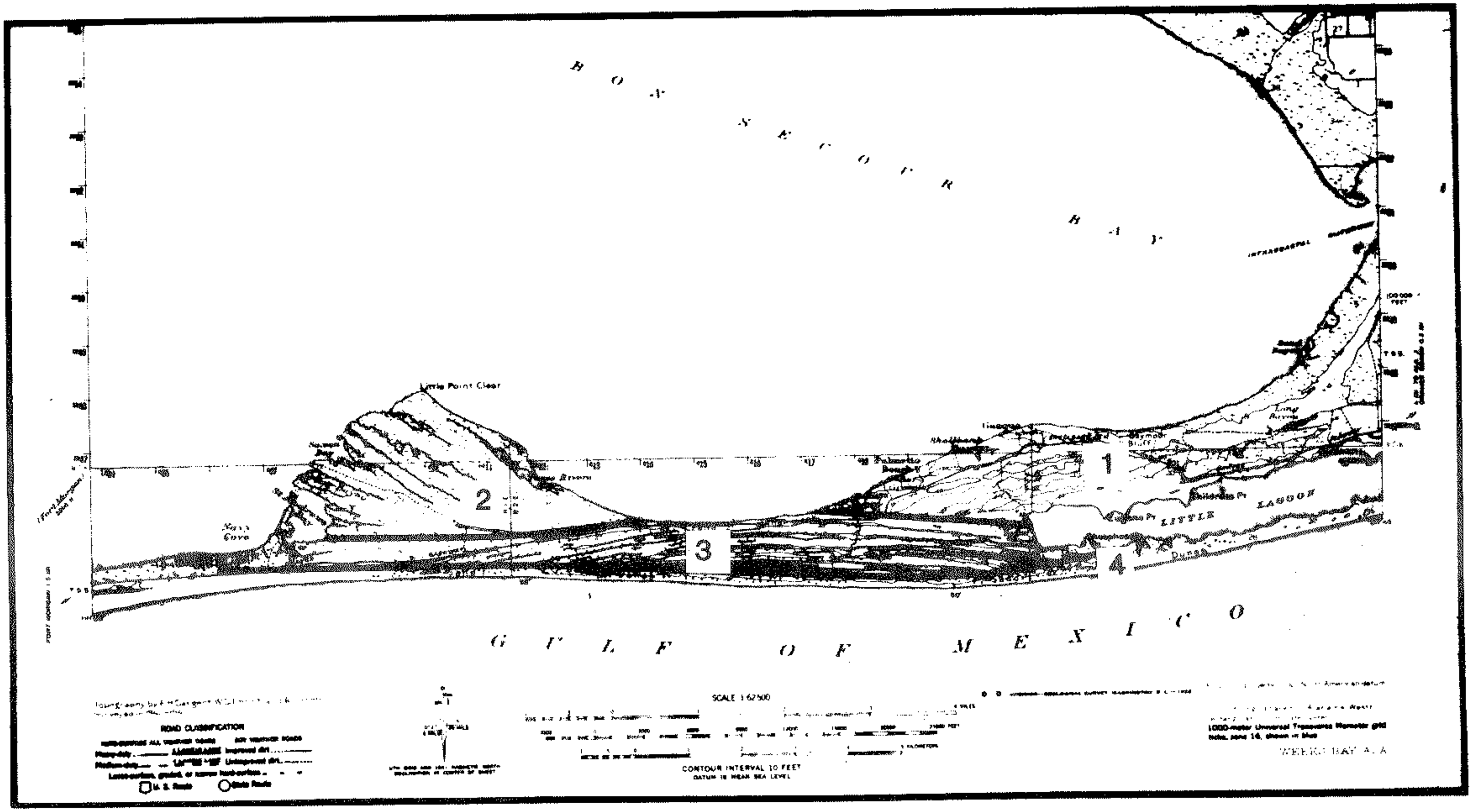

Figure 36. Topographic map of western Morgan peninsula showing (1) Gulfport Strandplain (late Pleistocene), (2) late Holocene beach ridge set submerged in saltmarsh, (3) younger ridge set, and (4) recent beach-ridge (dune complex, truncated earlier ridges, enclosed lagoon) (USGS Weeks Bay Topographic Quadrangle from Otvos 1985).

Figure 37. Aerial photograph of the western tip of Morgan Peninsula, showing Fort Morgan (A), Mobile Point, and Dixie Bar extending southward (Shepard and Wanless 1971).

The CBRS unit, however, does include a considerable amount of bay shoreline within five of the subunits.

Geologic History

The pre-Holocene land surface in the Mobile Point area is underlain by a transgressive-regressive sequence (Otvos 1979). Transgression during the Sangamon Interglacial (in the Pleistocene) resulted in the deposition of sandy muds and the development of a beach ridge along the western shore. Interbedded with and covering these are sedimentary deposits of the Prairie Formation. Lowering of sea level due to glaciation in late Wisconsinan time (late Pleistocene) resulted in river channel incision to about 100 ft below present sea level (Otvos 1985). Holocene transgression flooded Mobile Bay and shifted its shores to the present Mobile-Tensaw Delta region, about 30 mi north of today's bayhead. Buried oyster reefs at the head of the Bay probably relate to this stage, which preceded the southward progradation of the Mobile-Tensaw Delta (Ryan 1969). Presently, bay headwaters are too fresh for oyster growth.

As the marine transgression came to an end (about 3,000 years B.P.), the Holocene shoreline was established and strandplain progradation began. Multiple generations of beach ridges were formed on Morgan Peninsula, separated by truncated interfaces. The present straight shore sharply truncates the more indented earlier trends. The embayment in front of the original Perdido Bay entrance, located between two Pleistocene headlands, caused local drift reversal that prograded Ono Spit northeastward. Westward growth of the eastern segment of the most recent Perdido Key beach ridge completely isolated Big Lagoon from the gulf when its western end became attached to the Gulf Beach strandplain area. Late Holocene constrictions of the Mobile Bay entrance resulted from the growth of these ridge generations (Figure 36).

Modern Changes

Shoreline changes occur on both the gulf and Mobile Bay margins of the Morgan Peninsula. Erosion is a major concern along the Mobile Bay shoreline (Figure 38). Measurements of the change in shoreline configuration between 1917 and 1974 show that as much as 170 ft of erosion may have occurred from the mouth of Bon Secour River to Catlins Bayou during that time, although the average amount was closer to 50 ft (Hardin et al. 1976).

From Three Rivers to the eastern seawall of Fort Morgan, much erosion and shoreline modification has occurred (the CBRS unit is within this area). From Little Point Clear to St. Andrews Bay, bayward-projecting spits lost from 200 to 790 ft between 1917 and 1974. From Navy Cove to the eastern seawall of Fort Morgan, losses on the order of 200 ft were noted (Hardin et al. 1976).

Bathymetric data from 1929 to 1973 show that St. Andrew Bay, Navy Cove, and the bay north of Fort Morgan are becoming progressively shallower. The southwest cove of St. Andrew Bay has also become a shoal area. This trend probably reflects reworking and subsequent deposition of material eroded from

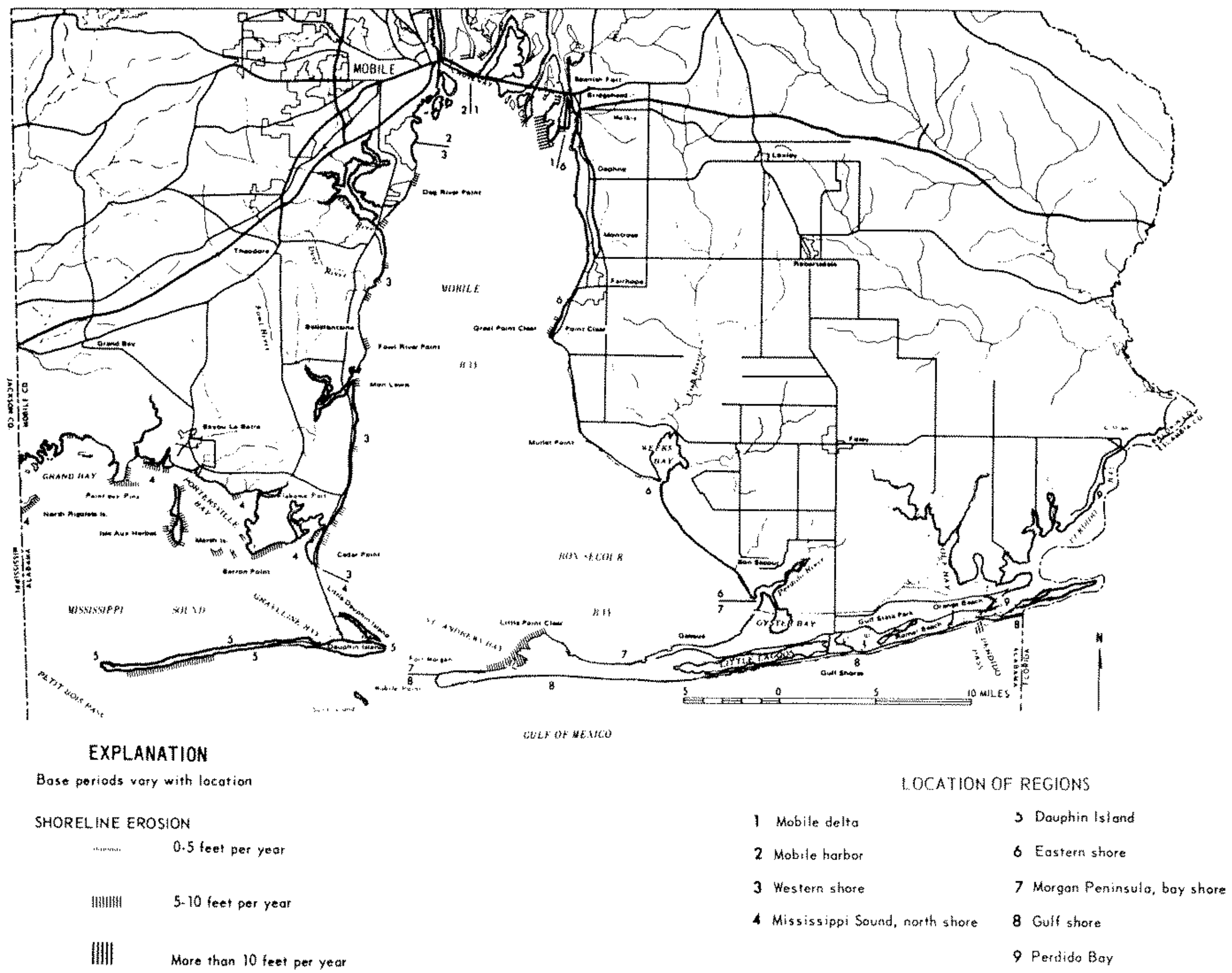

Figure 38. Net historical shoreline trends, Alabama coast, 1917-1974 (Hardin et al. 1976).

the shoreline west of Little Point Clear, material from the spoil banks on the south side of the Intracoastal Waterway north of Morgan Peninsula, and material from the spoil banks northwest of the peninsula along the Mobile Ship Channel (Hardin et al. 1976).

Variable shoreline trends are measured for the gulf shoreline of Alabama. The historical trend is toward accretion at Mobile Point, primarily due to westward transport of sediment by longshore drift. Presently, the depth of the tidal pass between the Morgan Peninsula and Dauphin Island (another unit within CBRS), which reaches 65 ft, prevents further westerly migration of the peninsula (Figure 39; Otvos 1985).

Eastward along the remainder of the Alabama coast to the Florida border, the shoreline eroded by an average of 80 ft from 1917 to 1974 (Hardin et al. 1976). In 1917, several tidal inlets existed, opening into Little Lagoon and Shelby Lakes. The inlet connecting Little Lagoon to the gulf was 0.7 mi west of Gulf Shores. This inlet was approximately 260 ft wide. A second inlet along the gulf shoreline was approximately 1 mi west of Romar Beach. This inlet connected the easternmost lagoon on the Shelby Lakes with the gulf, and was approximately 65 ft wide. By 1941, both of these inlets had closed and a second inlet to Little Lagoon had opened. This inlet, 2.2 mi west of the inlet of 1917, was about 200 ft wide. High-altitude infrared photographs taken of this area in 1974 showed no passes open into either Little Lagoon or Shelby Lakes, although some water possibly flows through the western inlet of Little Lagoon at the highest high tide (Hardin et al. 1976). At least one pass to Little Lagoon was opened by Hurricane Frederick (1979) but had closed by 1982 (CBRS photographs). Currently, a number of passes are open, cut by the storm surges of Hurricanes Elena, Juan, and Kate in 1985.

Coastal Change Processes

In coastal Alabama, average wind directions and velocities are seasonal. During the fall and winter months, winds are predominantly from the north or northwest, while spring and summer winds are from the south or southwest (Chermock et al. 1974). In January, north and northwest winds have average peak velocities of 8 to 11 mph. In June, south and southwest winds have average peak velocities of 8 to 17 mph.

Persistent high winds from the north and northwest during the winter months create wind tides which lower the water level in the northern part of Mobile Bay and build up waves along the south and southeast shores (including the CBRS unit). Under these conditions, severe erosion may occur along the northern shore of Morgan Peninsula. The situation is reversed during the summer and occasionally in the winter months. Waves and tides then build up on the upper bay, causing severe erosion along the western shore and lower Mobile Delta, and water and waves periodically cover the causeway across the lower Mobile delta. These variations in water level cause complex currents that complicate the circulation within the bay.

Coastal Alabama has experienced the effects of at least 32 tropical storms and hurricanes during this century, most recently Hurricane Frederick in 1979, and

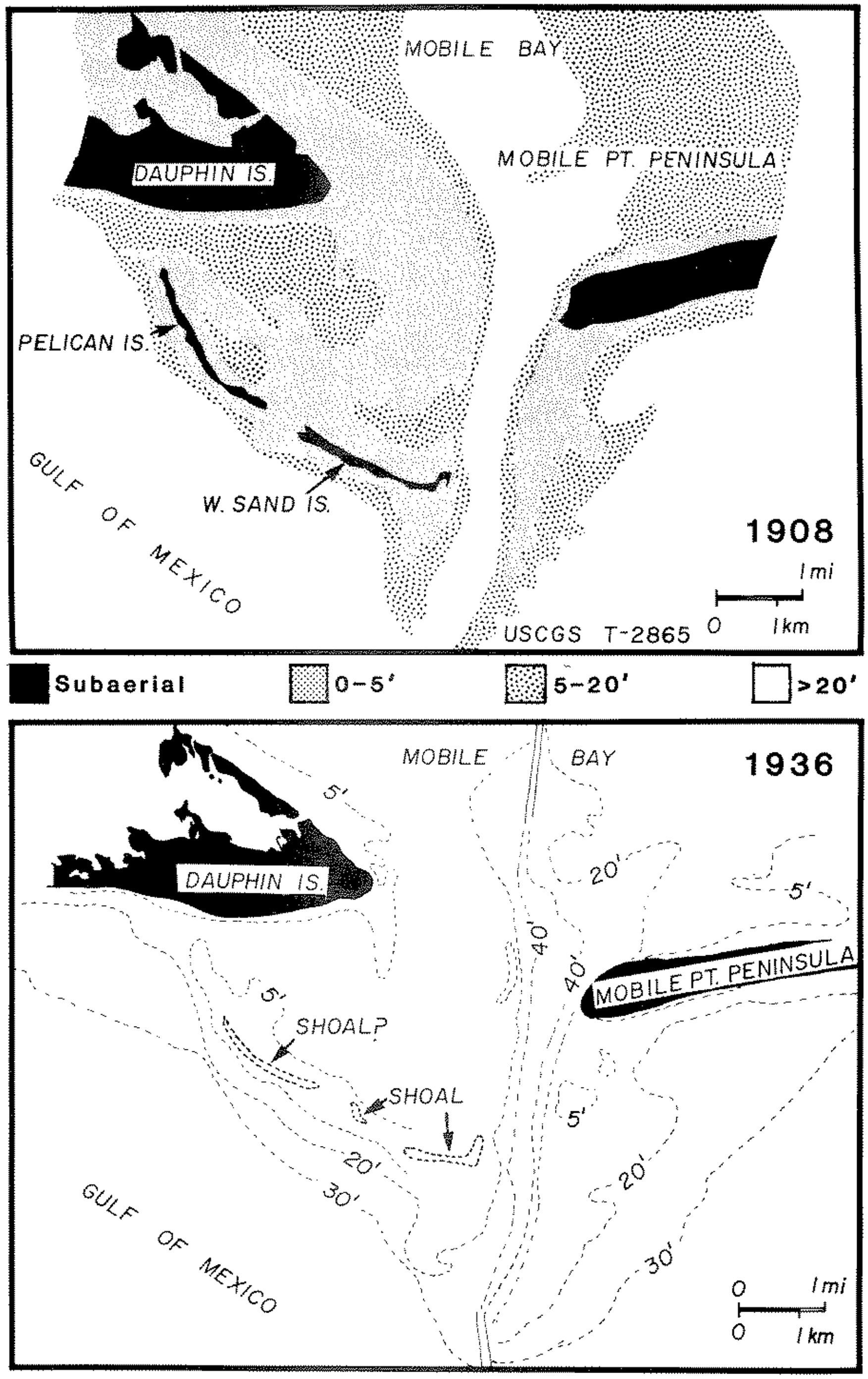

Figure 39. Shoreline configuration of Mobile Bay and associated passes, 1908-36 (from Otvos 1985).

Hurricanes Elena, Juan, and Kate in 1985 (updated from Hardin et al. 1976; Penland et al. 1980). Simpson and Lawrence (1971) calculated that for any 1 year the probability of landfall of a tropical storm in this area is 13%, of a hurricane 6%, and of a great hurricane 1%. The utility of such statistical treatments must be kept in perspective, as illustrated by the triple hurricane impact of 1985, following a 6-yr hiatus.

Sediment availability affects shoreline stability. Sediment is supplied to the Alabama coast by the Mobile River System and by the Perdido River system. Prevailing onshore, southerly winds produce westward longshore currents, which sweep sand along the Morgan Peninsula toward Mobile Point. The mouth of Mobile Bay serves as a major barrier to continued drift, resulting in much of the sediment being carried offshore by strong tidal currents.

Tide gauge analysis for Pensacola (to the east) indicates an apparent rise in sea level of 1.8 mm/yr over the past 45 years (Hicks 1978). This closely approximates the eustatic rise in sea level; thus, subsidence does not appear to be a major contributing factor to shoreline erosion in this region.

Human activities also affect coastal change processes. Erecting impermeable structures, mining sand for construction purposes, dredging and spoil disposal, and damming rivers can all affect the shoreline. Major impacts on the Mobile Point area and the Morgan Peninsula as a whole have come from the construction of Fort Morgan in 1819, from dredging and spoil disposal, and from the construction of some single family dwellings and condominiums along the Mobile Bay shore. As yet, none of these activities has had a significant effect on erosion of the gulf shoreline. However, shoreline erosion during the unusually active hurricane season of 1985 resulted in considerable dune truncation, placing structures in jeopardy.

Management Implications

Shoreline erosion will probably continue along both shores of most of the Morgan Peninsula, but the Mobile Point area (the geographic area--not the CBRS unit which is spread along much of the Morgan Peninsula area) will likely remain stable to accretionary. Because it is important in maintaining the estuarine character of Mobile Bay, care should be taken to preserve the integrity of the peninsula. Specifically, destruction and devegetation of foredunes should be avoided. Shore-perpendicular structures such as roads and driveways act as focal points for storm surge, thus accelerating breaching and overwash. Therefore, these should be minimized. Despite the relative stability of this area, evacuation routes are low-lying and easily flooded, so development should be kept to a minimum. Although Hurricane Frederick (1979) did extensive damage, the general region has undergone considerable growth following a 1- to 2-year hiatus immediately after Frederick. The triple hurricanes of 1985 will probably once again retard the growth rate for 1 to 2 years. This unit is possibly the best example of high development pressure and concomitant effects of hurricanes in the CBRS. The aerial photographs, which are available for dates before and after the hurricanes (Frederick and the 1985 hurricanes), and the insurance settlement claim data would provide material for an excellent, detailed cost analysis of development subsidies in a hazardous coastal area.

CBRS UNIT Q02--DAUPHIN ISLAND, ALABAMA

Geomorphology

Dauphin Island is a combination low- and high-profile barrier (Figure 40). The CBRS unit includes the low western end of the island and a small section on the bayside of the northeastern end of the island. The eastern end of the island consists of a Pleistocene core 3.1 mi long and 1.6 mi wide, with large landward-migrating sand dunes up to 45 ft high. This higher portion of the island is vegetated by a dense stand of pine. Prior to Hurricane Frederick (1979) the 12 mi long Holocene spit west of this core (the area in the CBRS unit) was characterized by an almost continuous washover terrace with small, discontinuous dunes. The barrier width ranged from 1,000 to 2,000 ft. The other barrier islands on the upper gulf coast--Petit Bois, Horn, and Ship Islands--are all high-profile regressive barriers.

Geologic History

The pre-Holocene land surface in the Dauphin Island area is similar to that described for the Mobile Peninsula area (CBRS unit Q01). At the future sites of Deer, Round, and eastern Dauphin Islands, the Pleistocene land surface rose slightly above the surrounding area. These sites are interpreted as parts of a large Pleistocene barrier complex along the northern Gulf of Mexico.

The high Pleistocene core of eastern Dauphin Island probably had a significant effect in capturing and shunting westward the sand that crossed the developing Mobile Bay ebb-tidal delta platform from Mobile Point Peninsula, which steadily prograded westward. The upper gulf coast islands became thinly veneered by beach and dune deposits, and their southern shores steered the littoral sand further west. Through time an extensive belt of shoals and emerging islands developed between eastern Dauphin Island and present-day New Orleans, Louisiana. The maximum age of Dauphin Island is about 3-4,000 years. This period coincides with the start of the relative stabilization of eustatic sea level.

Modern Changes

The shorelines of Dauphin Island have been greatly modified through its known history. Shortly after 1717, DuSualt, a Frenchman, produced a map of the island which indicated that at that date Dauphin Island and Petit Bois Island (presently immediately west of Dauphin Island) were connected (Hardin et al. 1976). The hurricane of 1740 breached this connection. This conclusion was reached because the "Isle Dauphine" shown on the circa 1717 map has a hump on the western spit very similar to the hump of the present day Petit Bois Island. Also, the next island to the west on the circa 1717 map was called "Isle a Corne" (Horn Island), which is the island to the west of the present-day Petit Bois Island.

U.S. Coast and Geodetic Survey Charts from the 1850's indicate that Dauphin Island was breached by an 1852 hurricane. Since 1900, Dauphin Island has been breached twice. Between 1909 and 1917 a hurricane divided the island into two smaller islands separated by 5.3 mi of open water, shoals, and scattered

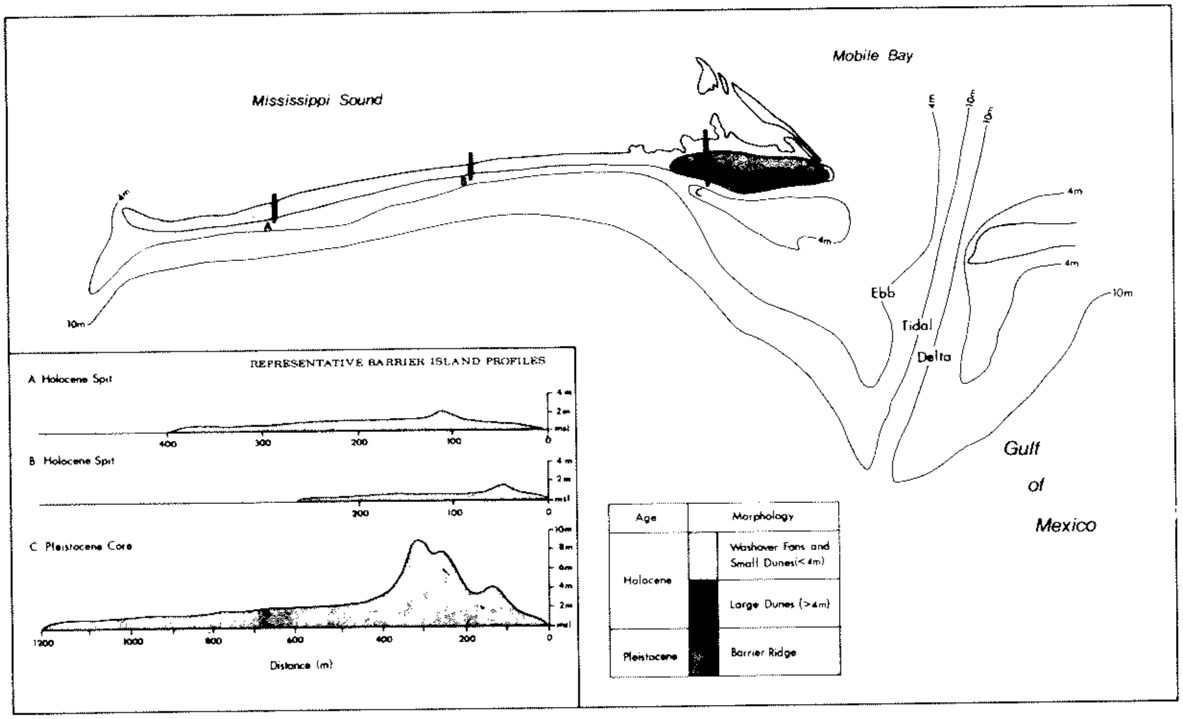

Figure 40. The morphology of Dauphin Island, Alabama (Penland et al. 1980).

remnants of the former island. The western island was 3.8 mi long and the eastern island (the Pleistocene core) was 4.1 mi long (Figure 41; Hardin et al. 1976). Another hurricane in 1923 prolonged the existence of this breach.

Between 1923 and 1942, the hurricane-created tidal inlet filled with sediment, thus rejoining the two parts to form one island. Air photos taken on March 23, 1950, show Dauphin Island again breached by the hurricane of September 4, 1948. Tides generated by this hurricane were reported to be 6 ft above normal (U.S. Army Corps of Engineers 1973). The island was breached about 4,000 ft west of Oro Point. The breached area was approximately 1,400 ft wide and, by the date of the photos, was probably covered only at high tide. A washover fan extended over much of the length of the island but was best developed at a distance of 2 mi west of Bayou Heron channel.

The barrier dune complex slowly migrated north as much as 1,600 ft between 1917 and 1942 (Figure 41). No measurable movement was detected between 1942 and 1958.

Figure 42 shows a general trend of erosion along the gulf shore of the island and general elongation of the western end of the island. Figure 43 shows the cumulative erosion over time at two locations along the gulf shore of the island. Shoreline erosion on the part of the island that was westernmost in 1917 totalled 580 ft from 1917 to 1974 or 10.1 ft/yr. Shoreline erosion on the entire gulf shore for the period 1942 to 1974 totalled 209 ft or 6.3 ft/yr excluding the accretion on the western tip of the island. This accretion added a total of 1.8 mi to the length of Dauphin Island from 1917 to 1974 (Figure 44; Hardin et al. 1976). In summary, the gulf shoreline of Dauphin Island has been undergoing fairly rapid erosion over the past 70 years. Material eroded from the shoreline is transported westward and is redeposited along the length of the island.

Persistent high winds from the north and northwest during the winter months tend to depress the water level in much of Dauphin Island Bay and concurrently cause a buildup of waves along the north shore of Dauphin Island, where wind fetch length is great. Under these conditions, severe erosion may occur along Dauphin Island's northern shore.

Coastal Change Processes

For discussion concerning seasonal wind speed and direction, hurricane probabilities, and apparent sea-level rise for the area, see discussion under Mobile Point, CBRS Unit Q01.

Management Implications

Extensive residential development has occurred on Dauphin Island since 1950. Because this island has been breached twice by hurricanes in this century and because hurricane frequency along this part of the gulf coast is high, there is every reason to believe that it will be breached again--at great cost to private property in the area. Paradoxically, most of the new residential development has occurred in areas most susceptible to storm damage, while

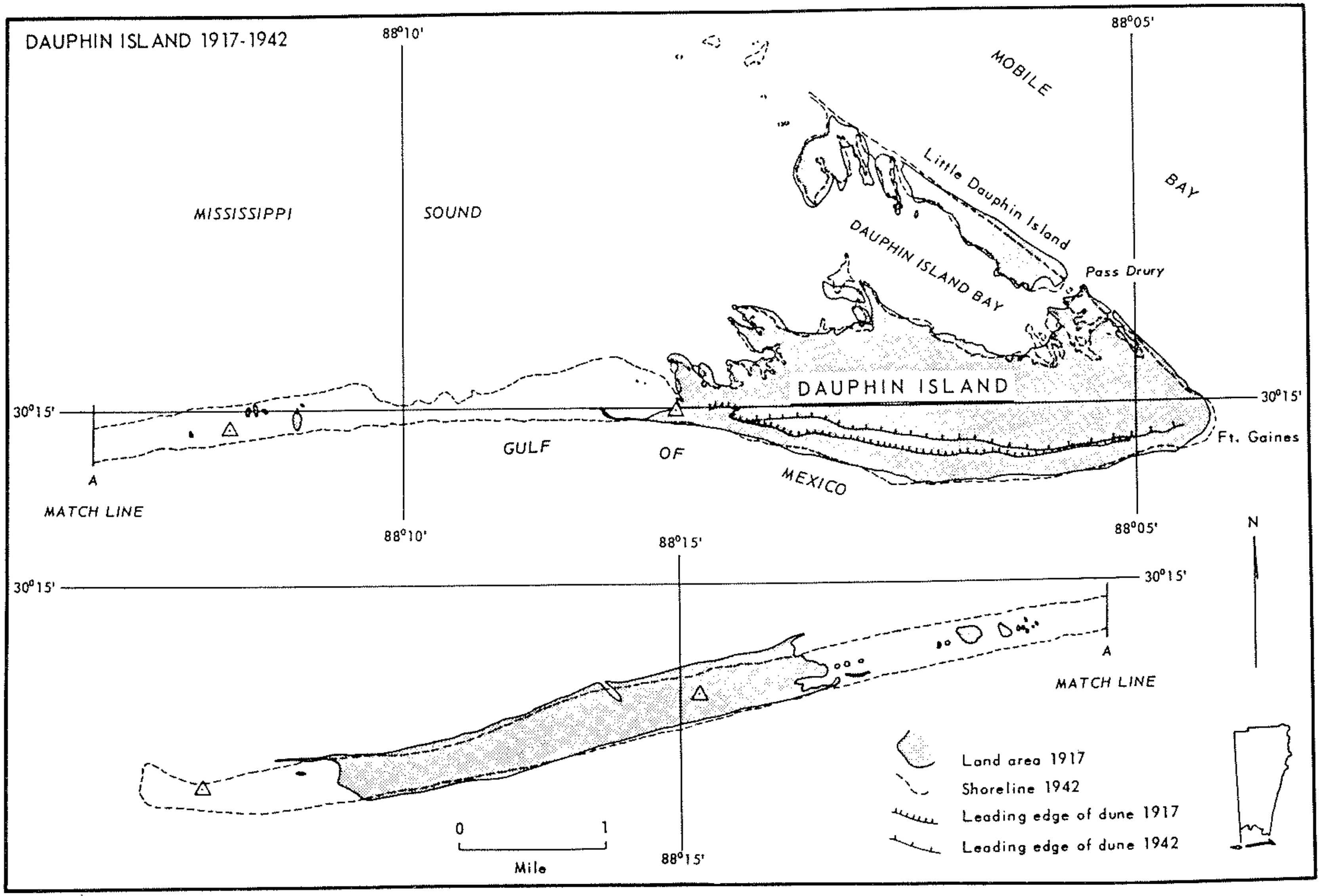

Figure 41. Shoreline change in Dauphin Island, Alabama, 1917-42 (Hardin et al. 1976).

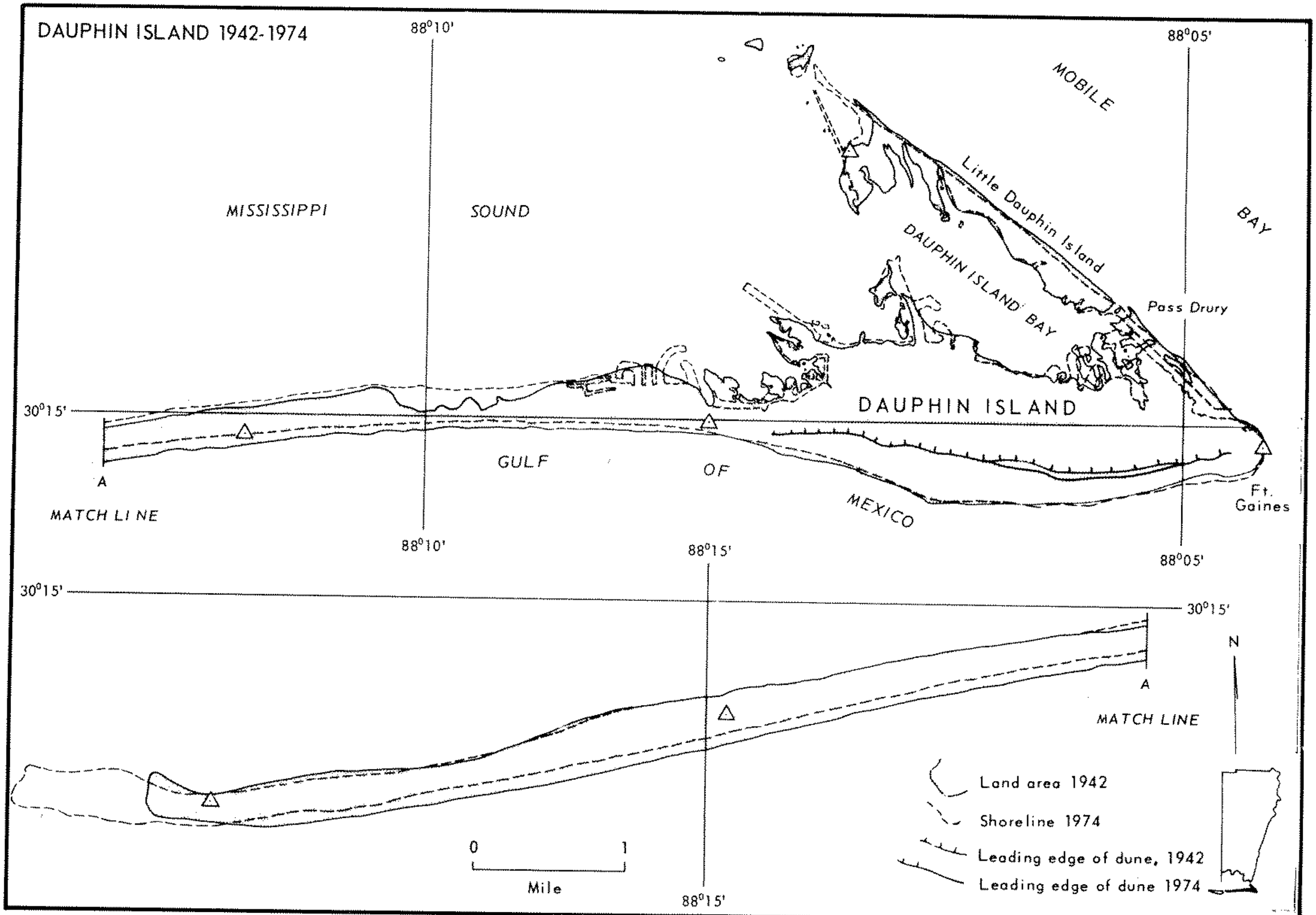

Figure 42. Shoreline change in Dauphin Island, Alabama, 1942-74 (Hardin et al. 1976).

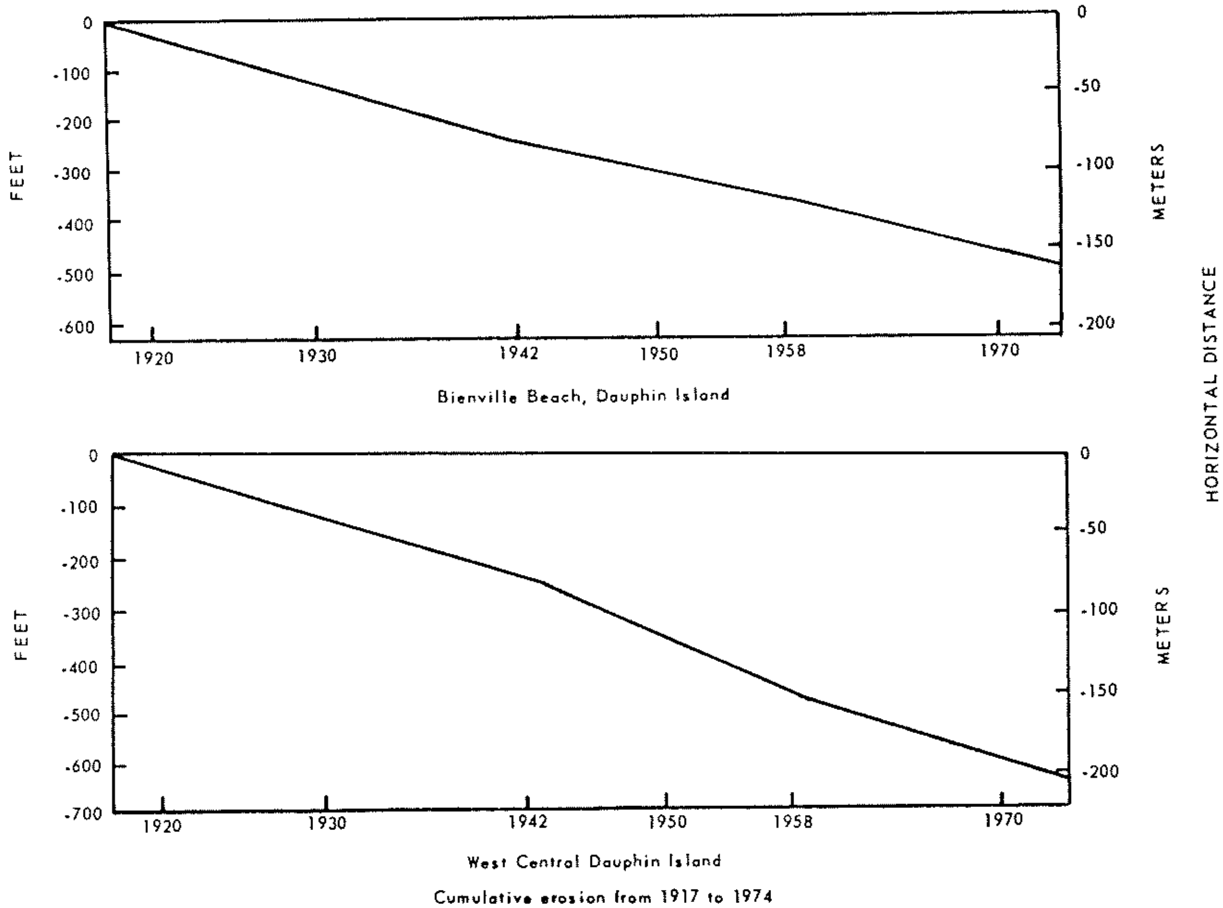

Figure 43. Cumulative erosion along Bienville Beach and west-central Dauphin Island, 1917-74 (Hardin et al. 1976).

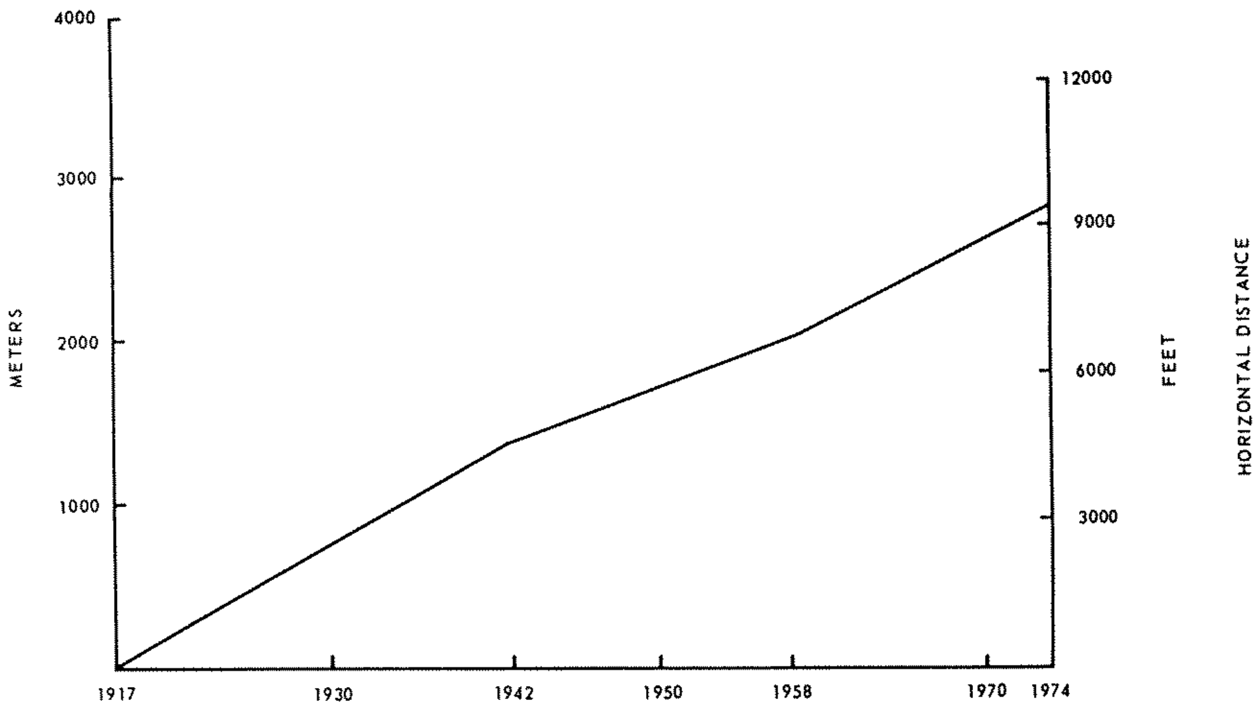

Figure 44. The westward accretion of the western spit of Dauphin Island (Hardin et al. 1976).

large tracts of subdivided land in the more stable and elevated eastern part of the island, which is protected by a large primary dune complex and a forest of pines, are relatively undeveloped.

Major efforts should be undertaken to predict the distribution of different hazard zones on the island prior to its further development. Development was poorly planned in the past as is illustrated by impacts of Hurricane Frederick which included the loss of the Dauphin Island Bridge--the only evacuation route (Penland et al. 1980).

CBRS UNIT S06--ISLES DERNIERES, LOUISIANA

Geomorphology

The Isles Dernieres are located along the southeastern coast of Louisiana (Figure 45). This symmetric barrier island arc is approximately 19 mi long. Typical barrier widths range from 0.9 to 1.2 mi in the central island arc to 0.3 to 0.6 mi in both lateral flanks. The Isles Dernieres comprise five islands separated by tidal inlets. These inlets are 1,000-4,000 ft in width, with depths ranging between 20 and 60 ft. Inlet morphology varies from wave-dominated to tide-dominated, depending on age of the barrier and the size of the tidal prism. In the east-central portion of this barrier island arc, the remnants of a beach ridge plain, the Cheniere Caillou, can be recognized.

The Isles Dernieres are extremely low-lying and highly dynamic. Elevations are generally less than 5 ft. Dune development is poor. The lee side of the islands contains salt marsh intermixed with black mangrove.

The Isles Dernieres barrier chain has a combination of characteristics that makes it different from the other CBRS case studies: (1) the development of this barrier system is distinctly different (see Geologic History); (2) the rate of shoreline change is extreme (see Modern Changes); (3) the entire land area of the island group and, more importantly, a substantial part of the sediment-sharing system are within the CBRS; and (4) the area has experienced relatively little human perturbation.

Geologic History

The Isles Dernieres represent one of three basic evolutionary stages of barrier system development within the Mississippi Deltaic Plain (Penland et al. 1981). The deltaic plain was formed from a series of overlapping delta systems deposited over the past several millenia (Figure 46). Each of these major delta systems had a lifespan of one to two thousand years. A delta system will continue to lengthen and enlarge until a more hydraulically efficient route to the coast is established. This is believed to occur by means of "stream piracy." After the main flow of water is split, usually during a flood, the newer, shorter, and steeper path to the sea progressively captures more of the river's flow and begins to build a new delta lobe. This process is currently going on in the Atchafalaya River as it attempts to capture more of the Mississippi River's flow.

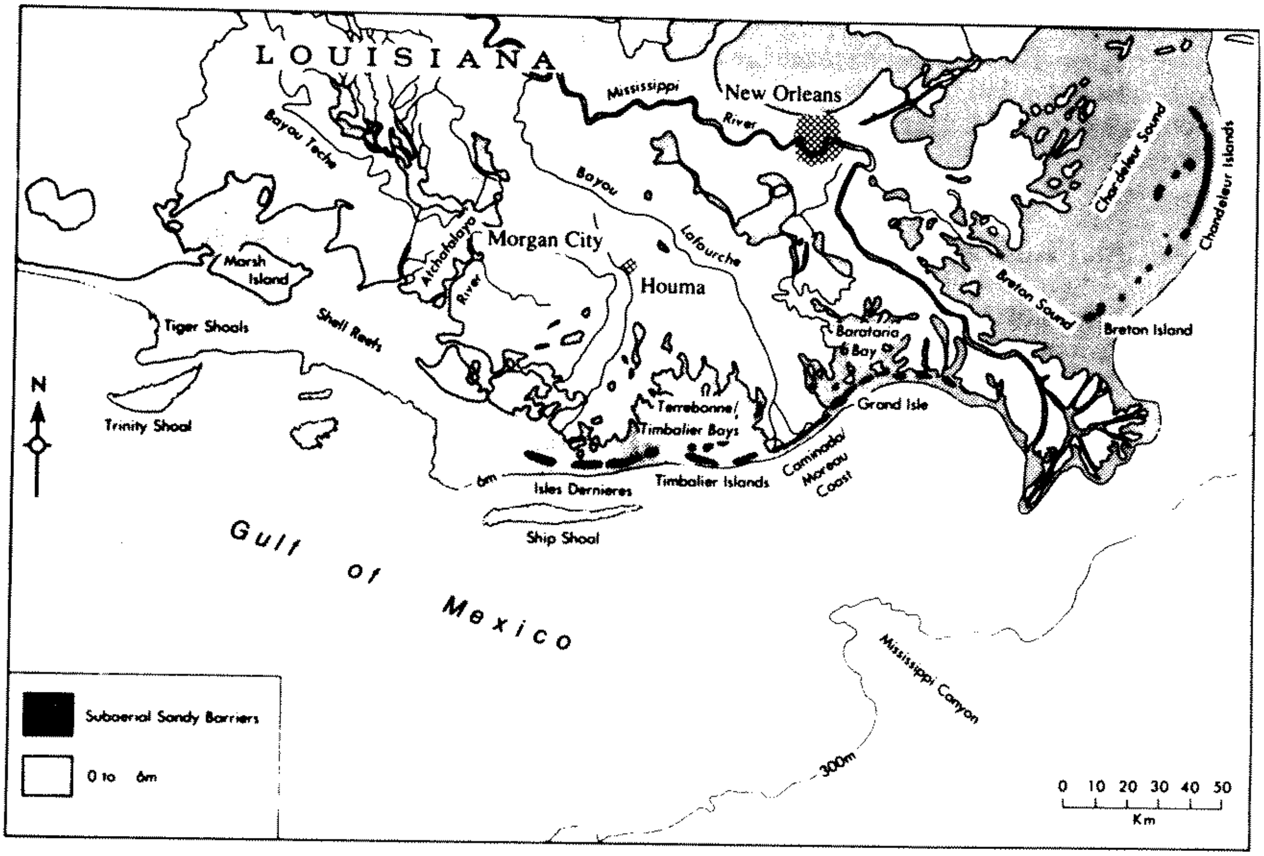

Figure 45. Location of Isles Dernieres along the south-central coast of Louisiana (Penland and Boyd 1982).

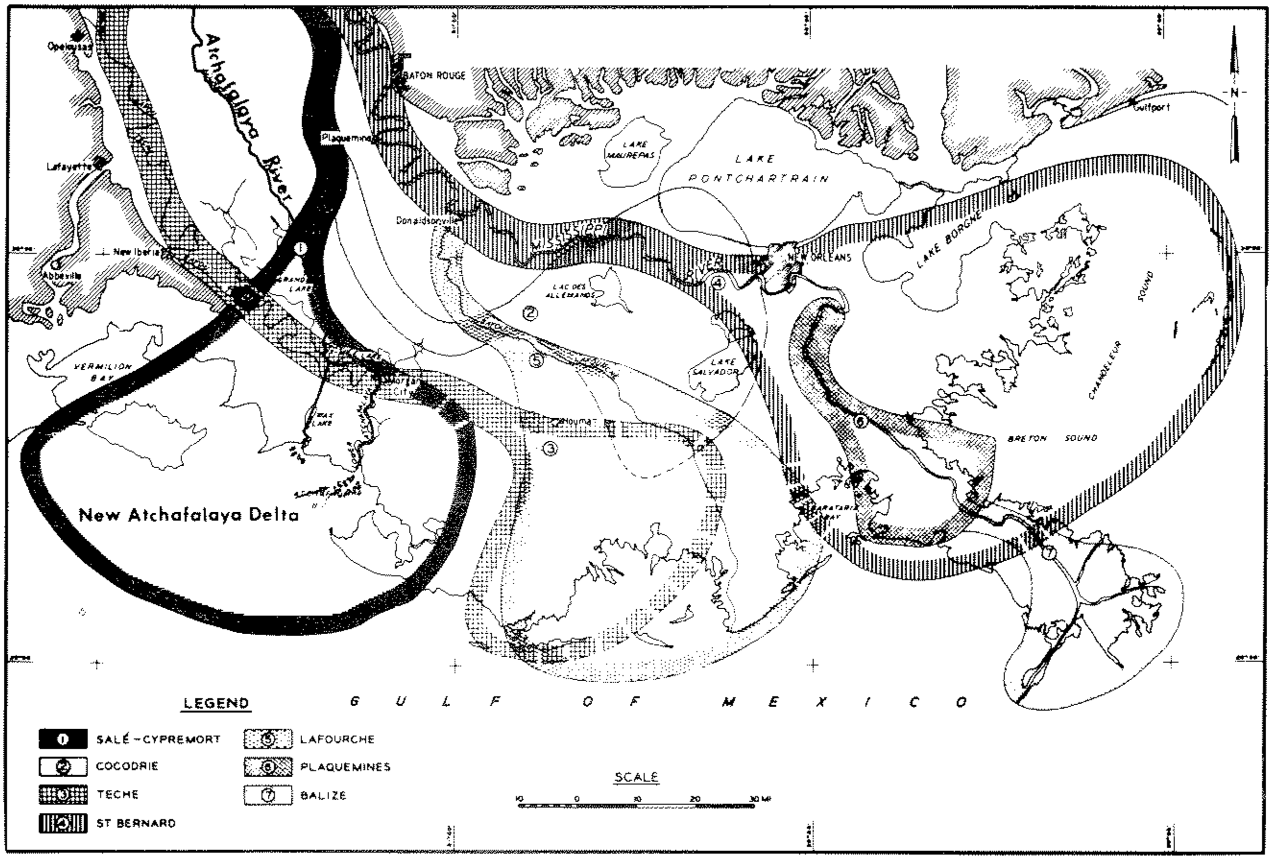

Figure 46. Major delta lobes of the Mississippi River Deltaic Plain (Roberts and Heerden 1982 as modified from Kolb and Van Lopik 1966).

The deterioration (destruction) phase of a delta system begins with trunk-stream abandonment. As new sediments are no longer supplied, erosional processes become dominant over progradational processes in the abandoned delta (Frazier 1967). It is within this latter part of the delta cycle that a cycle of sandy barrier island development and destruction occurs.

The abandoned delta system is first transformed into an erosional headland with flanking barrier spits and beaches (Penland et al. 1981). This stage of development is exemplified by the present mouth of Bayou Lafourche and extends east and west to the Grand Terre Islands and Timbalier Islands, respectively. Segments of the Grand Terre and Timbalier Islands are also part of the CBRS (Units S02 through S05) and are located immediately to the east of the Isles Dernieres.

Continued marine transgression leads to the second stage of barrier island development--the transgressive barrier island arc (Penland et al. 1981). The Isles Dernieres are representative of this stage and differ from the previous stage in that there is no longer an erosional headland to provide coarse clastics. As the delta lobe continues to subside, seawater encroaches behind the barrier and gradually the entire island system becomes completely separated from the mainland. The finite amount of sand left on the barrier continues to be reworked in response to waves, storms, and hurricanes until the third and final stage is reached. This final stage is characterized by the complete loss of any subaerial or emergent land; the barrier becomes a sandy inner-shelf shoal (Penland et al. 1981). Several of these shoals are present off south-central Louisiana and are associated with the oldest of the Mississippi Deltaic Plain delta complexes--Teche and Maringouin (4,000-6,000 years B.P.).

Modern Changes

The Isles Dernieres lie at the approximate seaward boundary of the Early Lafourche Delta, which was abandoned between 300 and 1,000 years B.P. (Morgan 1974). The earliest surveys of the area (Landreth 1819) indicate that the Isles Dernieres were at first a single island (Isle Dernier, also known as Last Island) narrowly separated from the mainland (Stage 1). By 1853, the island was separated from the mainland by Pelto and Big Pelto Bays. By 1974, these two bays had enlarged and coalesced to form Lake Pelto, and the island had separated into five islands which were some 7 km from the mainland (Penland and Boyd 1981; Figure 47).

Several investigators have measured the shoreline erosion and the decrease in areal extent of the Isles Dernieres over time (Morgan and Larimore 1957; Peyronnin 1962; Morgan 1974; Adams et al. 1978; Penland and Boyd 1981). The Isles Dernieres face directly into the dominant southerly wave approach. This results in high erosion rates, particularly during periods of hurricanes (Penland and Boyd 1981). Since 1932, the islands have experienced an average annual shoreline retreat of some 24 ft with erosion rates in excess of 49 ft/yr common along the central portions of the island arc (Morgan 1974; Adams et al. 1978). Erosion rates are lower along the eastern and western extensions of the island system because material eroded from the central Isles Dernieres continues to be transported to the flanks, slowing the apparent

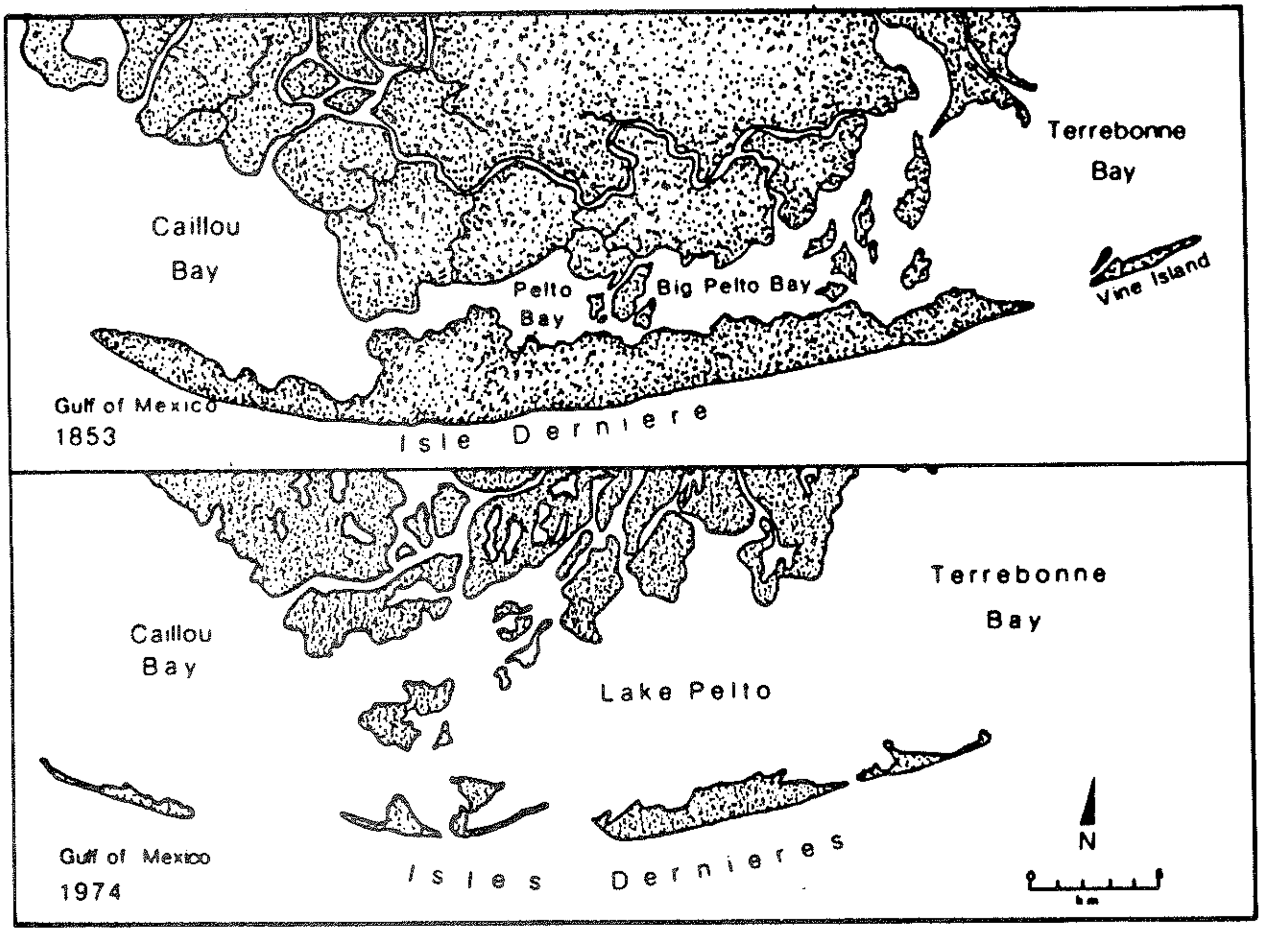

Figure 47. Comparison of Isles Dernieres shoreline in 1853 and 1974 showing detachment of the barrier from the mainland and the breakup of the land mass (Penland and Boyd 1982).

erosion measured there. Since 1853, the Isles Dernieres have retreated more than 0.6 mi landward and the areal extent of the islands has decreased from 14.8 mi^2 in 1887 to 3.9 mi^2 in 1979 (Penland and Boyd 1981). Today, the Isles Dernieres have narrow beaches with limited elevation and are frequently overwashed. The marsh areas on the sound side of the islands are rapidly eroding. The subaerial portions of the islands are expected to disappear and transform into subaqueous shoals (Stage 3) in 50 to 100 years (Peyronnin 1962; Penland and Boyd 1981).

Coastal Change Processes

The northern gulf coast is a microtidal, storm-dominated environment. Nearshore energy levels resulting from wind and wave processes are low except for the winter passage of cold fronts and the summer occurrence of hurricanes and lesser tropical storms.

Tides are diurnal with a mean range of 1.2 ft. Frequently, water level and water flux are wind dominated. Cold fronts (10-25 per winter season) will elevate and depress mean sea level between 1 and 4 ft (Boyd and Penland 1981). Storm surges associated with hurricanes range from 6 to 23 ft above mean sea level (Boyd and Penland 1981). Hurricanes have a probability of occurrence of 12% in any given year (Simpson and Lawrence 1971) and tropical storms have a recurrence interval of 1.6 years (Boyd and Penland 1981). Maximum storm surge levels can overwash the entire CBRS unit, flooding the coast for many miles inland. Hurricane Juan (1985), after stalling off the coast to the southwest of Isles Dernieres and placing the CBRS unit in the quadrant of highest winds for several days, completely leveled the islands (Figure 48). Investigations show the entire island chain retreated between 35 and 100 ft during this hurricane (S. Penland, Louisiana Geological Survey; pers. comm.).

Apparent sea-level rise for this region is an important factor, which affects the whole Mississippi Deltaic Plain system. Tide gauge data analysis from areas to the east, west, and north of the CBRS unit depict a trend of rising sea level (approximately 1.2 cm/yr) during the past three decades (by far the highest rate for all the CBRS case studies). Eustatic sea-level rise has been estimated to be 0.12 cm/yr (Gornitz et al. 1982), thus accounting for only 10% of apparent sea-level rise in this area. The remaining 90% is attributed to subsidence.

Management Implications

The extremely rapid disintegration of the Isles Dernieres makes it highly unlikely that any attempts at permanent development will occur. Historically, however, Isles Dernieres was a resort area. The hurricane of August 10, 1856, destroyed the large hotel Trade Wind (also reported as Muggah's Hotel) and 100 cottages and claimed approximately 200 lives (Aquanotes 1973). The resort was never rebuilt. There is some oil and gas activity on and around the islands and canalization has contributed to the land loss there, especially in the central portion.

Management strategies should be largely passive. Because the island system is naturally deteriorating at such a rapid rate, engineering projects to retard

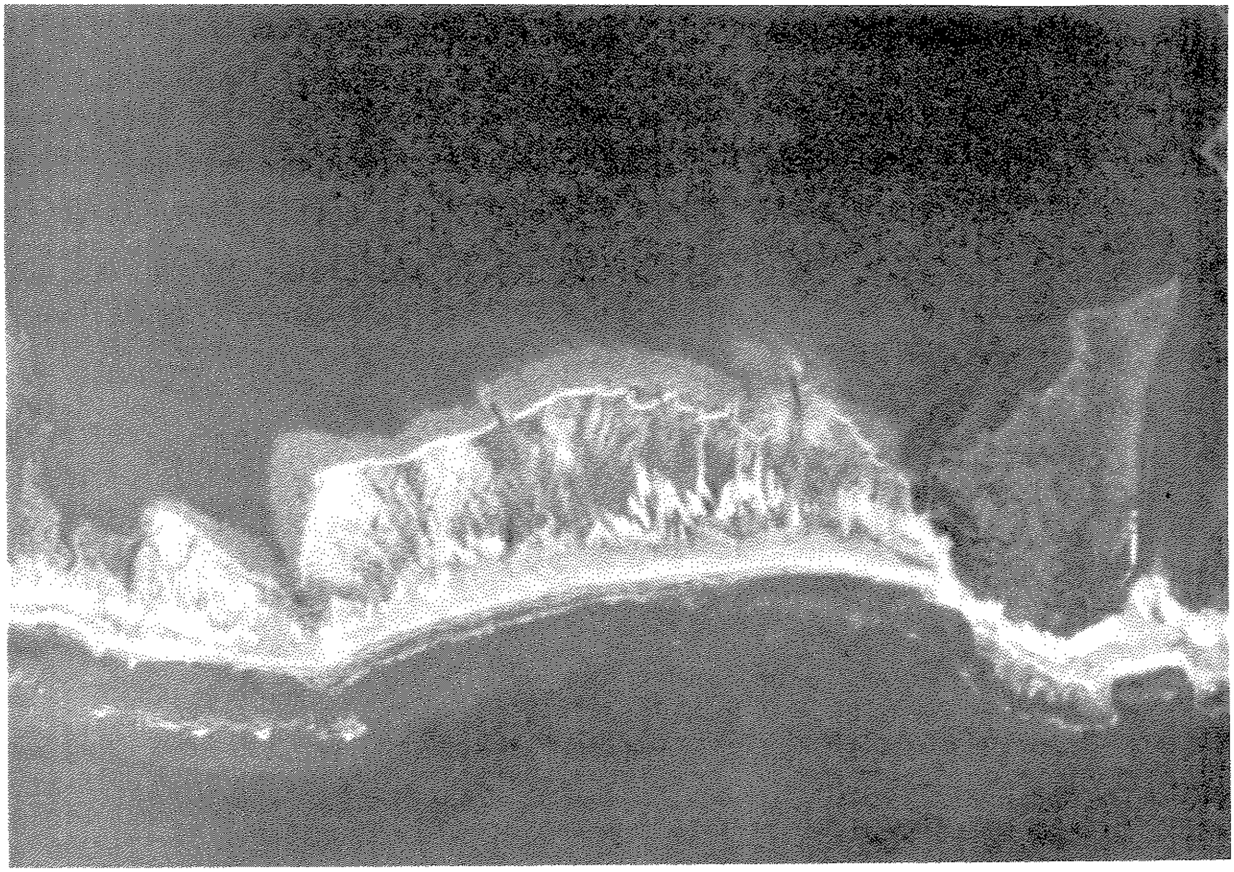

Figure 48. Aerial photograph of a section of Isles Dernieres after Hurricane Juan, October 1985 (S. Penland).

erosion are not likely to have a high benefit to cost ratio, as has been previously concluded by the U.S. Army Corps of Engineers (Peyronnin 1962).

The island arc has considerable fish and wildlife habitat value. The nearshore and bay waters contain sandy bottoms which are relatively uncommon along the Louisiana coast. This habitat is extensively used by both commercial and recreational fishermen and has been since at least the early 19th century (Landreth 1819) despite the relative remoteness of the area. The land area of the islands serves as an important bird rookery (Keller et al. 1984) although the continued decrease in dune development due to storms (Boyd and Penland 1981) will probably decrease the islands' nesting habitat value. Low-cost attempts to enhance dune development, such as planting stabilizing vegetation or using sand fences, would help to maintain this habitat, although we must realize that this habitat will probably still disappear within the next century through natural processes.

The island arc also serves the obvious ecological function of a barrier. Without the Isles Dernieres to absorb storm energies, the landward marshes both behind the islands and attached to the mainland would be eroding at an accelerated rate. Thus, continued protection of the CBRS unit itself serves to protect a wider area of valuable wetland as well. For this reason, and others, the State of Louisiana has planned a $16 million project to pump sediment onto the Isles Dernieres. The addition of sediment is designed to lengthen the existence of the Isles Dernieres in recognition of their function as a storm buffer to inland marshes. Eight of the CBRS units in Louisiana (S01-S08) are excluded from the CBRA prohibitions against federal funding of shoreline stabilization projects.

CBRS UNIT T03A--BOLIVAR PENINSULA, TEXAS

Geomorphology

The Bolivar Peninsula is the northernmost member of a chain of barrier shorelines that make up the gulf coast of Texas (Figure 49). Both low and high profile barriers occur on the Texas Coast. The transgressive low-profile barriers are found in association with deltaic headlands, and the regressive, high-profile barriers are located in the inter-deltaic bights (White et al. 1978; Morton 1979). (See Volume 1 of the report for a more detailed description of barrier types.) Galveston Island and the Bolivar Peninsula are regressive, high-profile barriers that lie between the Brazos-Colorado and Trinity Deltaic headlands (Morton 1979). The Bolivar Peninsula probably formed by westward spit accretion from the late Pleistocene Trinity Delta (Fisher et al. 1972). Today, an artificial inlet at Rollover Pass separates the peninsula from its headland to the northeast. The CBRS unit extends discontinuously from Rollover Pass westward to near the terminus of the spit.

A transect of the peninsula from the gulf to Galveston Bay shows a relatively wide beach (150 ft) with a well-developed, densely vegetated beach ridge and swale system containing ridge heights locally reaching 10 ft; a vegetated barrier flat; and salt marsh environments. Total width of the barrier averages in excess of 1.1 mi. The fan-shaped bulge on the back-barrier near

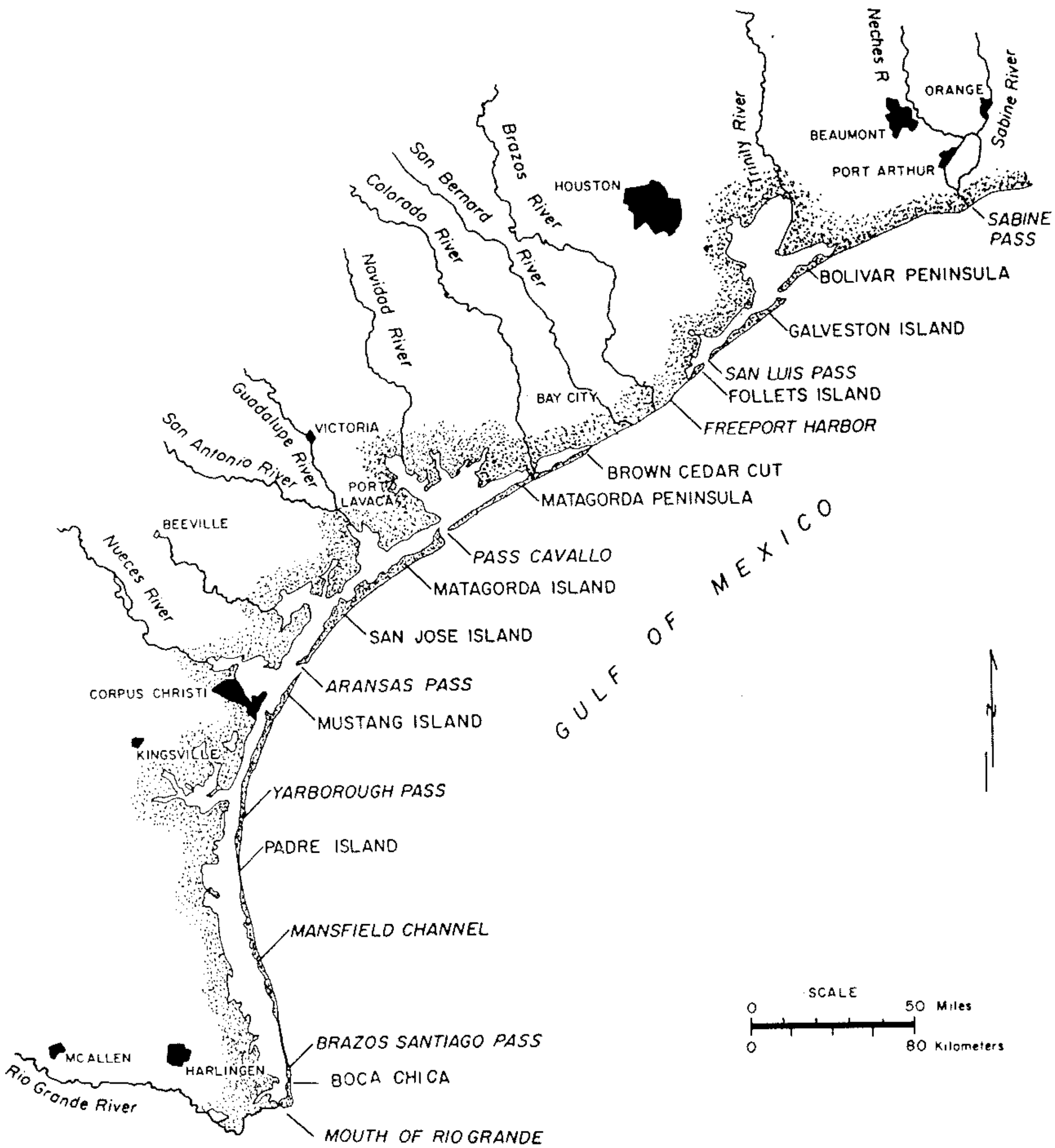

Figure 49. Map of the Texas coastline showing the location of Bolivar Peninsula, Follets Island, and Boca Chica (Morton and McGowen 1980).

Crystal Beach is a relict flood-tidal delta (Fisher et al. 1972). This morphology, together with curving beach ridges and inlet fill sequences in the subsurface, attests to the former existence of tidal inlets (Morton 1975; Morton and McGowen 1980). Subsurface borings reveal that barrier sands exceed 20 ft in thickness, increasing from east to west (Figure 50).

Geologic History

As the Holocene transgression began, about 18,000 years B.P., the Trinity River began filling in its valley. The modern Trinity River Delta developed as sea level reached its present stand some 3,000 to 3,500 years B.P. (Morton 1979). At this time, East Bay developed as an elongate lagoon behind the Bolivar Peninsula, which grew southwestward by spit accretion and onshore transport of sand from the eroding deltaic headlands (Trinity River System). The prominent ridge and swale topography of the western peninsula shows that seaward accretion was an important process (Morton 1975). Storm washover deposits and tidal deltas began filling the lagoons and estuaries on the bay side. Marshes were established and Bolivar Peninsula took on an aspect similar to that of the present day. Eventually, depletion of offshore sand sources and diminished supply of coarse sediment from the rivers combined to alter the conditions that led to seaward accretion (Fisher et al. 1972).

Modern Changes

Shoreline changes on the Bolivar Peninsula during recorded history have been mainly a result of human activities and storm impact. Earliest available records are from 1851 (Morton 1975). From that time until the Galveston jetties at Bolivar Roads Inlet were completed, the western end of the Bolivar Peninsula experienced up to 660 ft of erosion, probably due to repeated storm impacts in 1854, 1867, 1875, 1879, and 1900. The 1900 storm was a major hurricane that killed over 6,000 people on Galveston Island and was responsible for initiating the construction of the Galveston Seawall.

Construction of the jetties at Bolivar Roads, begun in 1874 and completed in 1915, and the deepening of the Bolivar Roads channel altered the pattern of coastal processes (Morton 1975). A large amount of sediment was eroded off the ebb-tidal delta of the pass and deposited on both sides of the jetties, with most of this sand accreting on the Galveston Island side (Morton 1977). Sand supplied from the eroding ebb-tidal delta, along with interception of longshore drift by the north jetty, enabled the extreme west end of Bolivar Peninsula to accrete some 1,000 ft seaward by 1930 (Morton 1975) despite the impacts of numerous hurricanes (Figure 51).

In 1955, construction of the artificial Rollover Pass was completed. The pass has not had a great deal of effect on shoreline erosion in western Bolivar Peninsula, which has been stable to slightly accretionary (Morton 1975; Figure 51). Since 1955, erosion immediately to the west of Rollover Pass has been about 10 ft/yr and decreases westward. Slight accretion is occurring in the immediate vicinity of the jetties (Morton 1975). Future trends will reflect the interaction of storms and the decreasing supply of longshore sediments from the eroding updrift sources. Hurricane impacts in recent times include 1957, 1961 (Carla, during which most of the peninsula was inundated), 1963,

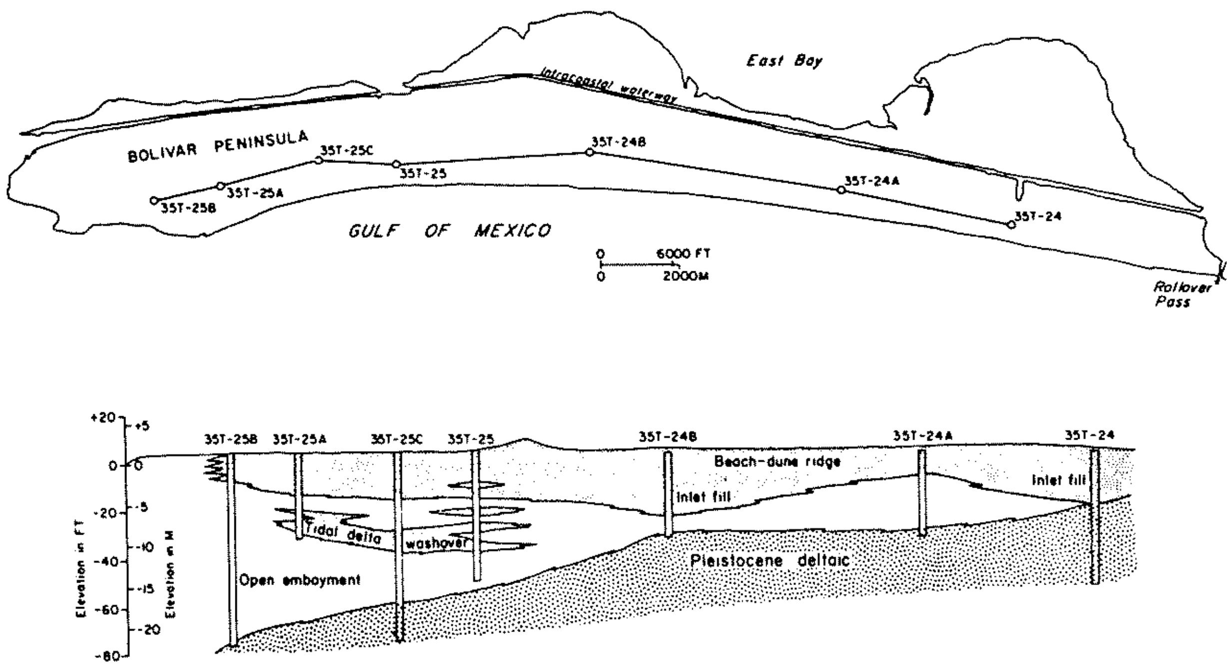

Figure 50. Geologic cross-section along the Bolivar Peninsula (Morton and McGowen 1980).

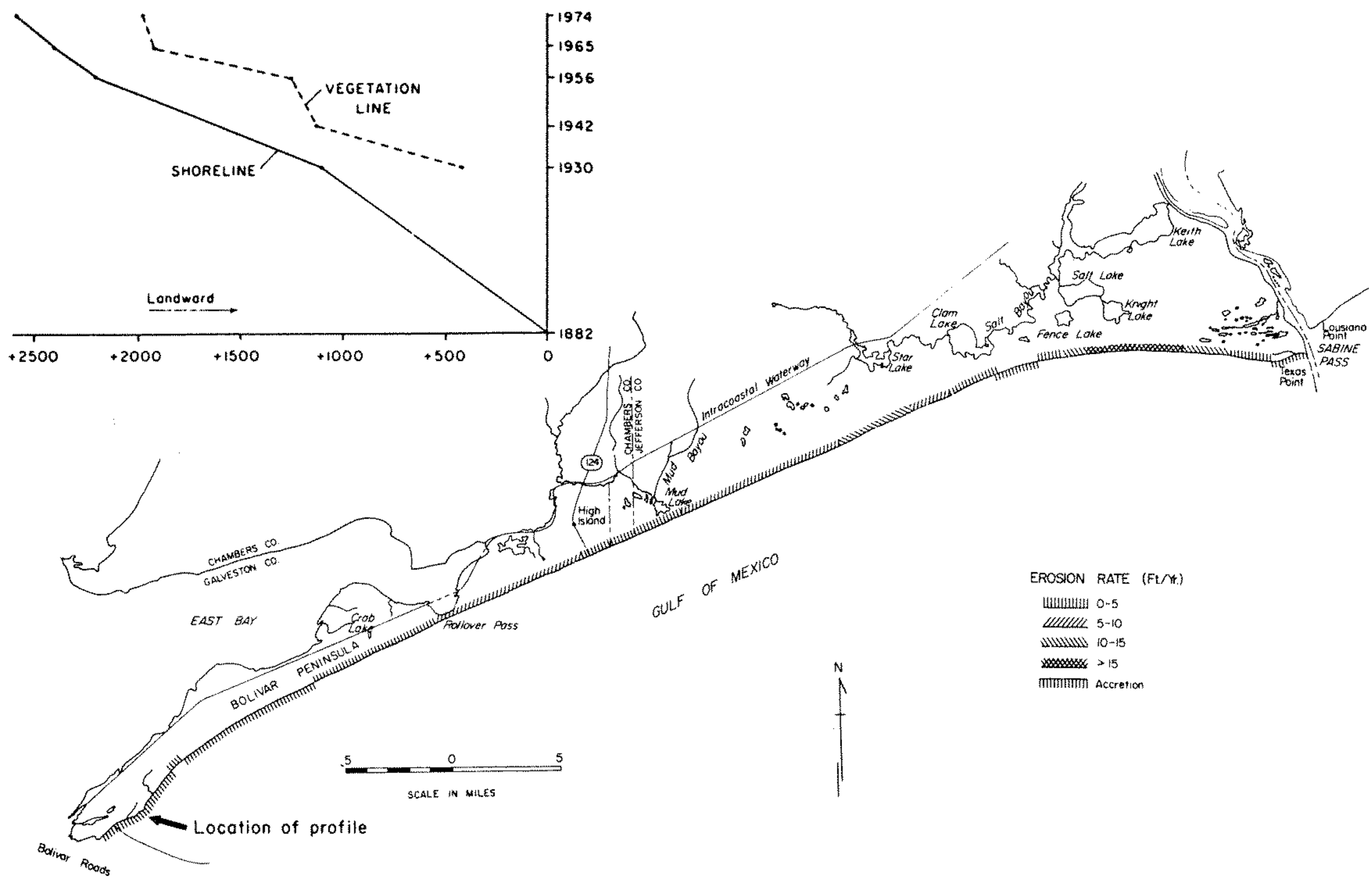

Figure 51. Net historical shoreline changes along the Bolivar Peninsula, and changes in the shoreline and vegetation line at a selected site indicated by the arrow (Morton 1975).

1970, 1973, and, most recently, Hurricane Alicia in 1983. Alicia's surge flooded parts of the peninsula, but the principal geologic effect was the cutting of ebb channels, which allowed ponded surge waters to return to the gulf.

Coastal Change Processes

Rainfall is an important factor in determining coastal change along the coast of Texas, as periods of dune migration have corresponded to drought years (Price 1958). Along the north Texas coast, precipitation usually exceeds evapotranspiration by 5-8 inches/yr (Fisher et al. 1972). The result is relatively dense vegetation and minimal aeolian transport of sediments. Thus, the wind regime has its greatest effect on this area by generating waves, which drive longshore currents, and not by forming coastal dunes. Moderate southeast winds prevail from spring to fall, but strong northerly winds accompany polar cold front passages in winter months (Morton and McGowen 1980). The dual wind systems result in seasonally variable directions of sediment transport, although net drift is to the southwest.

Normal astronomical tides are less than 3.3 ft, as compared to recorded open gulf surges of 13.3 ft (Morton 1975). Effects of a given hurricane vary depending upon its characteristics--size, central pressure index, duration, speed, and angle of approach--and characteristics of the area impacted, including shoreline and Continental Shelf configuration, and the stage of the astronomic tide. The relatively uniform height of the continuous foredune ridge along the Bolivar Peninsula decreases the probability of overwash breaching, and erosional retreat of the dune line is the more likely occurrence of a direct storm impact.

Another important factor controlling shoreline change is sediment supply. Sands which built the Bolivar Peninsula were derived from shelf sediments (McGowen et al. 1977), the Pleistocene Trinity Delta, and perhaps the Mississippi River (Morton 1975, 1979). The shelf sediment supply, more important in earlier stages, is now largely depleted (Morton 1975). Sand supply to the Bolivar Peninsula is now mostly limited to updrift erosion of its eastern end and the Trinity Deltaic headland.

Sea-level rise is important because a minor rise in sea level can cause considerable landward displacement of the coastline (Morton 1979). Eustatic sea-level rise has been documented (Gornitz et al. 1982), but is subordinate to apparent sea-level rise caused by compactional subsidence of deltaic sediments in this area (Swanson and Thurlow 1973; Morton 1979). Groundwater withdrawal for municipal and industrial use and withdrawal of hydrocarbons have also caused subsidence and faulting in this general area (Brown et al. 1974; Kreitler 1977); these exacerbate natural subsidence and true sea-level rise. Surface faulting is visible on the washover fan near the northeastern boundary of the CBRS unit (Ewing 1985). Subsidence, probably resulting from deep fluid withdrawal from the Caplen oil field, has resulted in the flooding of much of this fan since 1930 (Ewing 1985). Other human activities that have resulted in coastal change on the Bolivar Peninsula include the jetty construction at Bolivar Roads and the construction of houses and finger canals for boat access. These developed areas may serve as foci for future storm surges and increasing erosion. Although the CBRS unit excludes these already developed areas, it is naturally linked to these areas of development and impacts in developed areas will therefore extend to the CBRS unit.

Management Implications

The Bolivar Peninsula is already partially developed and has been episodically subjected to hurricane-induced inundation. A primary concern is evacuation routes. The ferry between Port Bolivar and Galveston will not operate under high wave conditions and Texas Highway 87 along the peninsula is at a low elevation, with a high potential for flooding. Thus, the evacuation routes of the peninsula can be easily cut off.

Care should be taken to prevent devegetation and destruction of foredunes. Shore-perpendicular structures in the developed areas should be kept to a minimum because they can serve to concentrate storm surge. The boundaries of this CBRS unit are discontinuous and checkerboard because they omit areas containing pre-CBRA development. This lack of geographic continuity could present management difficulties.

CBRS UNIT T04--FOLLETS ISLAND, TEXAS

Geomorphology

Most of the features that typify low-profile barriers are found on Follets Island. Follets Island is actually a peninsula located on the northern flank of the Brazos-Colorado Deltaic headland (Figure 49). A transect of the island from gulf to bay shorelines shows a narrow (<3,000 ft), low (<5 ft) barrier composed of beach, discontinuous dunes, vegetated barrier flat, and salt marsh with associated tidal creek environments (Morton 1982). The beach is predominantly fine sand and usually about 100 ft wide. Hummocky, discontinuous dunes 3 to 6 ft high occur. The large separation between these dunes leads to sheet overwash rather than channel and fan washover deposition (Suter et al. 1982), although some channeling does occur. Coalescing washover deposits make up the barrier flat, which merges with extensive back-barrier salt marsh (Morton 1982).

Most of Follets Island is part of the CBRS. Developed areas at the northeastern (San Luis Pass) and southwestern ends (Swan Lake vicinity) are excluded. In addition, there are three small discontinuations of the CBRS due to pre-CBRA development.

Geologic History

The development of Follets Island is tied to that of the Brazos-Colorado Delta. Approximately 18,000 years B.P. when sea level began to rise worldwide (Curray 1960), the Brazos-Colorado Delta was located much further seaward, near the Continental Shelf margin (Winker 1982; Suter and Berryhill 1985). Sea level rose rapidly but intermittently, until reaching a position near that of today about 3,000 to 3,500 years B.P. At this time the Brazos-Colorado

River began filling in the estuarine system in its river valley and a delta began to prograde seaward (Fisher et al. 1972). Neither the time nor location of the maximum seaward extent of the Brazos-Colorado Delta is known, but evidence suggests it was seaward of its present location and it occurred sometime between 1,800 and 1,000 years B.P. (period of abandonment of the Caney Creek course of the Colorado River) (Fisher et al. 1972; McGowan et al. 1976; Morton 1979). The reworking of coarse-grained deltaic deposits by marine processes, together with spit accretion caused by longshore movement of sediment on the northern flank of the deltaic headland, provided the necessary sediment for the development of Follets Island (Morton 1979).

Modern Changes

Follets Island has been retreating landward for over a century. Accurate coastal maps extend back to 1852. The northern end of Follets Island was once known as San Luis Island, separated from the remainder of the barrier by a tidal inlet known as Cold Pass. This inlet was open on the 1852 and 1867 coastal maps, but had closed by 1930 (Morton and Pieper 1975). Everywhere else on Follets Island, erosion from 1852 to 1930 averaged about 16.5 ft per year. In 1881 jetties were constructed at the Old Brazos River mouth (Oyster Creek), resulting in some 1,960 ft of progradation along the southern end of the CBRS unit (Morton and Pieper 1975).

In 1929, the diversion channel to the New Brazos River mouth was constructed, shifting active coastal deposition southward beyond the boundary of this CBRS unit. Erosion increased along the length of Follets Island to an average of over 20 ft/yr, and was greatest near San Luis Pass and least near the jetties at Freeport Channel (the old Brazos River mouth). In 1941 and 1942, hurricanes reopened Cold Pass (Morton and Pieper 1975).

Erosion of the old Brazos delta area (CBRS Unit T05, Brazos River) supplied some sediment to Follets Island and accretion began to occur near San Luis Pass. Significant erosion resulted in 1961 from Hurricane Carla, which brought a storm surge of close to 13 ft, inundating the entire island (Fisher et al. 1972). Depletion of the sediment supply has resumed the erosional trend, so that the net changes along the barrier have been highly erosional (Morton and Pieper 1975; Figure 52). The impact of Hurricane Alicia in 1983 caused significant channeling through the barrier, as well as considerable erosion near San Luis Pass (Dupre 1985).

Coastal Change Processes

The dominant processes resulting in coastal change of Follets Island are similar to those previously discussed for the Bolivar Peninsula (T03A). The human activities that have had the greatest effects on Follet's Island are the jetty construction on the Brazos River in 1881-1908, the diversion of the Brazos in 1929, and the periodic dredging of Freeport Harbor (Morton and Pieper 1975). The few houses which exist probably have little effect on shoreline change, except to destroy dunes and vegetation, but the access canals to some of these dwellings can serve as foci for overwash breaching during hurricanes.

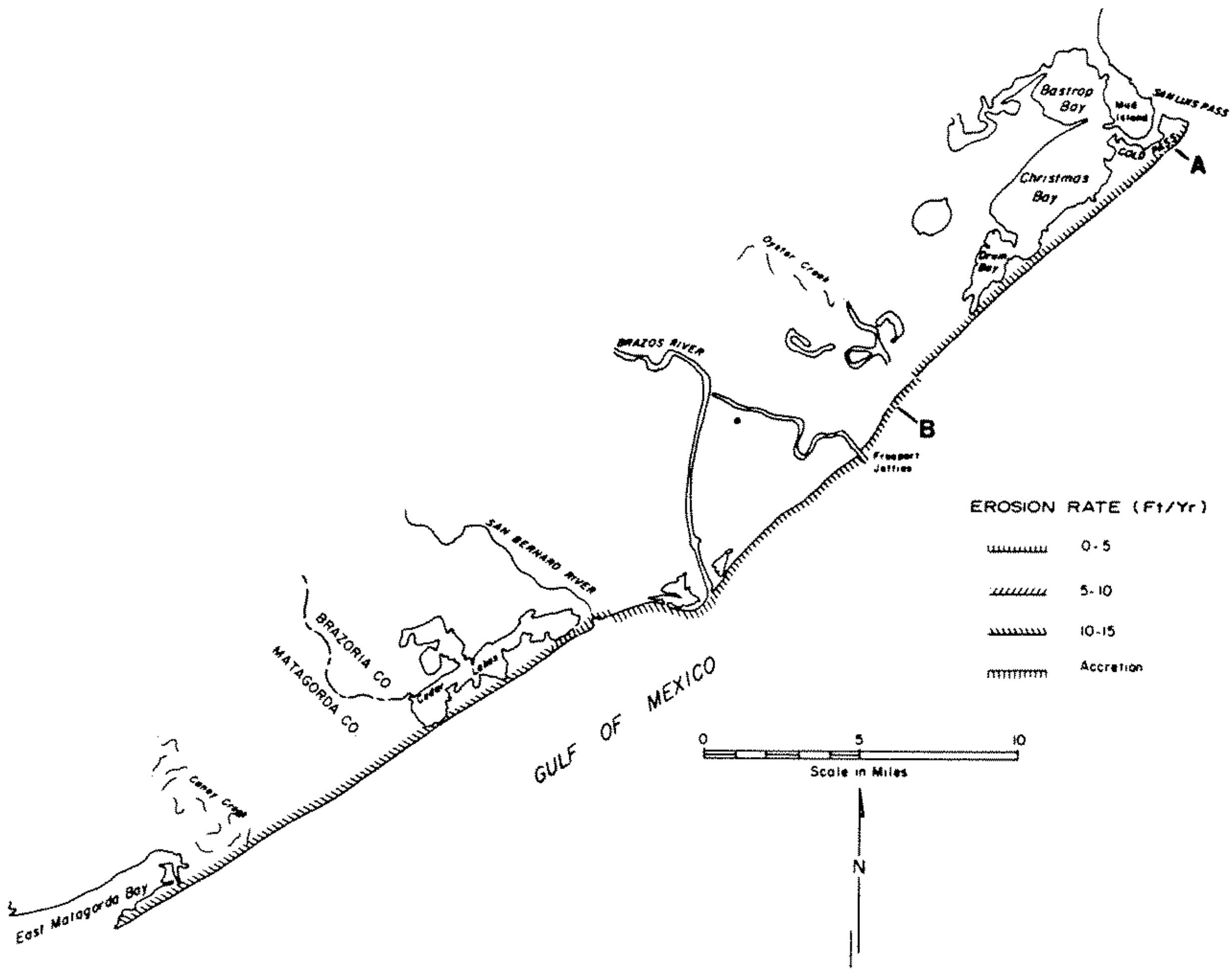

Figure 52. Net historical shoreline changes (1852/56-1974) along Follets Island and nearby areas (Morton and Peiper 1975).

Management Implications

Follets Island will probably continue to erode landward at rates of 6-10 ft per year (Morton and Pieper 1975). The entire island is considered a hazardous site for development because of the potential for flooding by hurricanes and the single evacuation route, which can be easily flooded and overwashed. The eastern end of the island at San Luis Pass is especially vulnerable to rapid shoreline changes. Thus, development throughout the unit should be discouraged, and erection of coastal structures avoided. Development pressure is present despite the rapidity of coastal change. In fact, the area immediately adjacent San Luis Pass (not within the CBRS), which has historically been subjected to the greatest amount of coastal change within Follets Island, is developed. This unit has a high fish and wildlife habitat value. It contains important nursery grounds for fish and shellfish species of commercial and recreational value.

CBRS UNIT T12--BOCA CHICA, TEXAS

Geomorphology

The Boca Chica ("little mouth" in colloquial Spanish) area of Texas is located at the southern end of the State near the Mexican border (Figure 49). Brazos Island, on which Boca Chica is situated, is actually a segment of the Holocene Rio Grande Delta, defined by the mouth of the Rio Grande River to the south and the tidal inlet at Brazos Santiago Pass to the north. The central portion of the gulf side of the island (about 20% of the area) was until recently the Brazos Island State Recreation Area. Most of the remaining land area is part of this CBRS unit.

Geomorphic features of the area are illustrated in Figure 53. The gulf beach was constructed by transgressive reworking of coarse deposits within subsiding lobes of the Rio Grande Delta (Brown et al. 1980). The beach, composed of relatively fine sand (Shideler and Smith 1984), reaches up to 330 ft in width at the northern end of the CBRS unit, and ranges from 5 to 20 ft in thickness (McGowen et al. 1977; Brown et al. 1980). Foredune development is variable; well-vegetated "dune islands" up to 33 ft in height occur but there is no continuous foredune ridge. Extensive wind-tidal flats and washover deposits make up most of the back-barrier. Also present are large clay-sand dunes (McGowen et al. 1977). Boca Chica itself is a relict tidal inlet which once served as a tidal exchange pass for South Bay and is now open only following hurricane impact.

Geologic History

About 18,000 years B.P., the Rio Grande Delta and the south Texas shoreline lay near the edge of the Continental Shelf (Berryhill and Trippet 1980; Suter and Berryhill 1985). Rising sea level caused the transgression of the Rio Grande Valley and deltaic platform, resulting in progressive infilling of the valley (Brown et al. 1980).

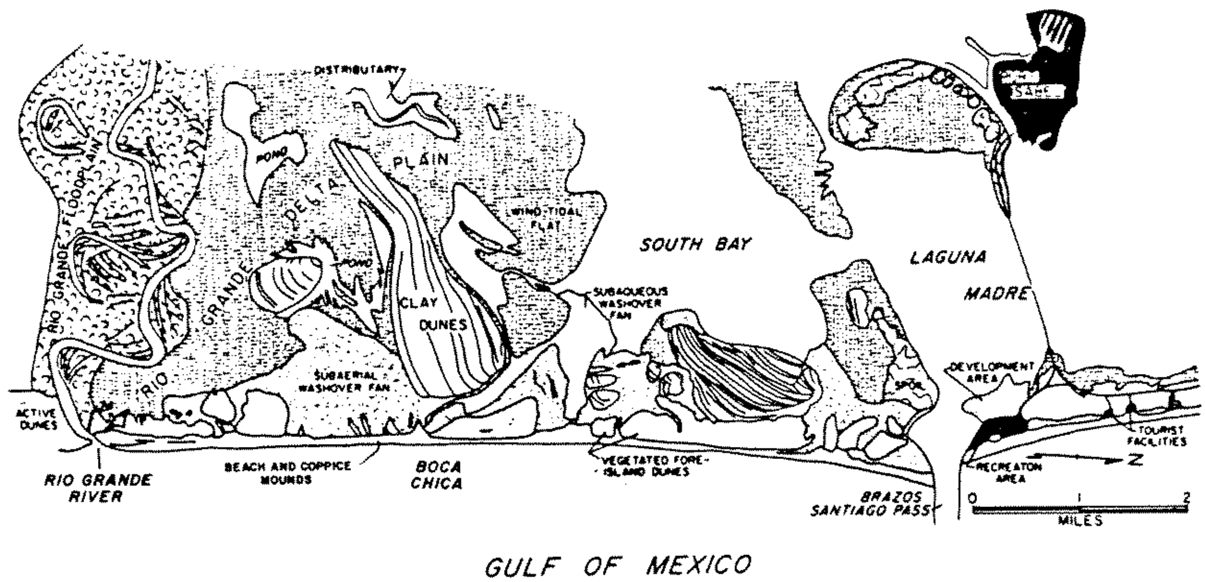

Figure 53. Geomorphology of the Boca Chica area (McGowen et al. 1977).

Some 7,000 to 10,000 years B.P. the Rio Grande Delta began to prograde seaward over earlier Holocene transgressive deposits (Fulton 1976). At least three lobes were deposited, extending perhaps 9 to 15 mi seaward of the present shoreline (Brown et al. 1980). Diminished sediment supply resulted in a lessening of delta growth as sea level reached its approximate current position some 3,500 years B.P. Subsidence of the delta lobes resulted in transgression, erosion, and the formation of South Padre Island between 3,400 and 1,900 years B.P. (McGowen et al. 1977). Brazos Island presumably formed around this time as well.

Aeolian processes and repeated hurricane impacts have caused a landward movement of the shoreline, shifting large amounts of sand into the Laguna Madre and South Bay. Predominant southeast winds created large foredunes, but the semi-arid climate prevents vegetation from stabilizing these features into a continuous foredune ridge. Wind tidal flats, resulting from meteorological tides caused by polar storm passage, began filling in much of South Bay and Laguna Madre. Ongoing compactional subsidence, transgression, storm impacts, and aeolian processes will eventually fill in all of South Bay and the southern Laguna Madre.

Modern Changes

Significant changes have occurred in the Boca Chica area during the last few hundred years, both naturally and as a result of human activity. Natural processes have been responsible for rather dramatic and repeated changes in the position of the mouth of the Rio Grande River (Figure 54). From 1854 to 1958 the river mouth migrated northward some 5,000 ft, then abruptly shifted some 4,000 ft south as a result of Hurricane Carla (1961). Hurricane Beulah (1967) interrupted a resumed northerly migration, and today the mouth of the river is in essentially the same position as in 1854 (Morton and Pieper 1975).

Brazos Island itself has experienced both erosion and accretion in contrast to South Padre Island (CBRS Unit T11) which has been retreating since the late 1800's. As a result of the construction of the jetties at Brazos Santiago Pass in 1935, the extreme northern end of Brazos Island is accretionary. The entire island has undergone net accretion since 1854 (Figure 55); however, since 1937, the island has been eroding at rates of 10 to 40 ft. Near the mouth of the Rio Grande River, erosion rates can be extreme (Morton and Pieper 1975; Brown et al. 1980). Construction of dams along the Rio Grande River, as well as diversion of water for irrigation, have essentially terminated the delivery of sediment down the Rio Grande and increased the erosion problem (McGowen et al. 1977; Brown et al. 1980).

Coastal Change Processes

Factors affecting coastal change are a complex interaction of natural and human causes. The climate of South Texas is semi-arid (Morton and McGowen 1980), although the area is subject to great variability in rainfall (Price 1958). During periodic drought years, sparse vegetation offers less resistance to wave and wind attack, thus increasing erosion. Similarly, severe freezes during intense winter storms, such as happened in 1983 and

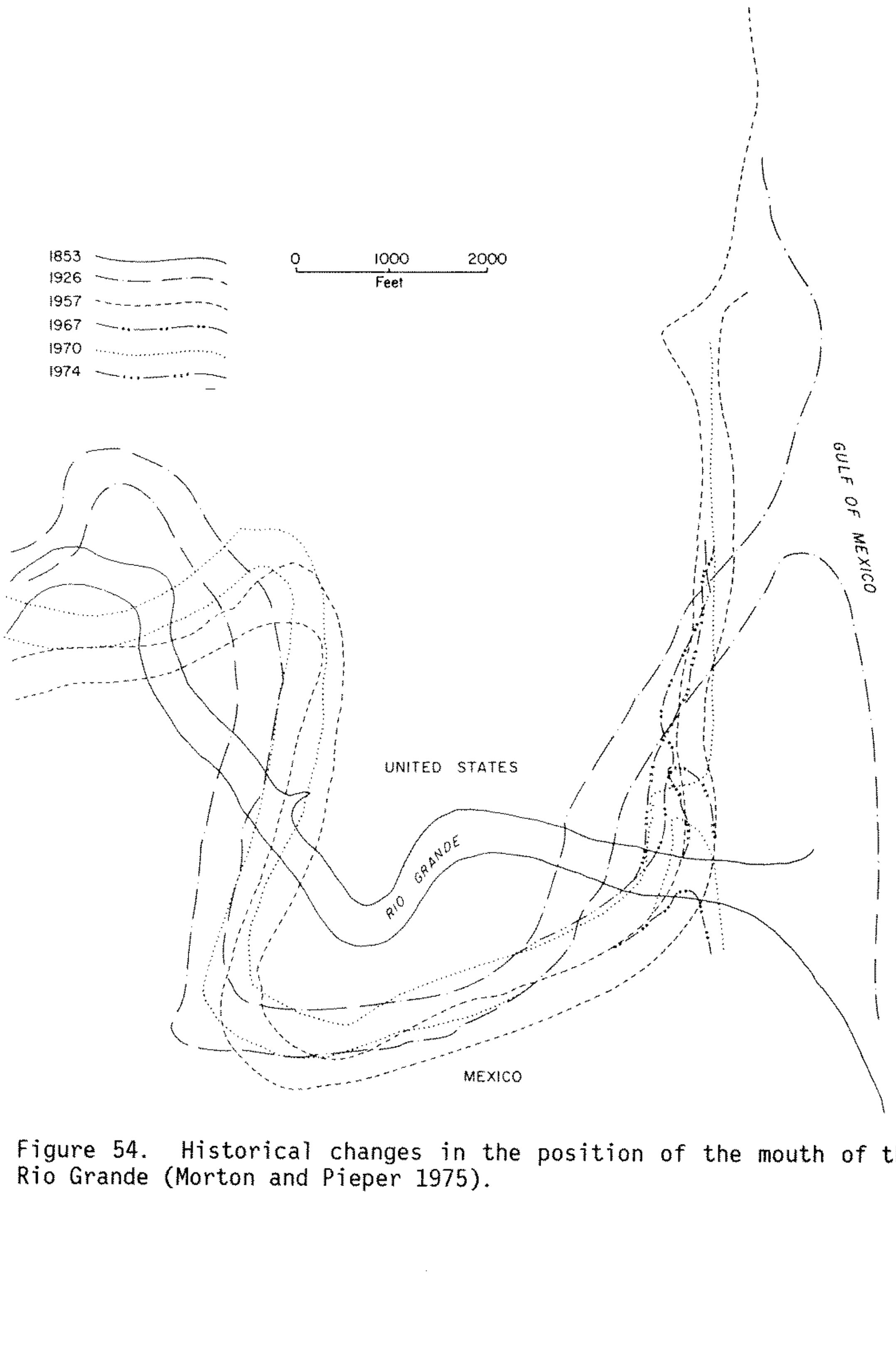

Figure 54. Historical changes in the position of the mouth of the Rio Grande (Morton and Pieper 1975).

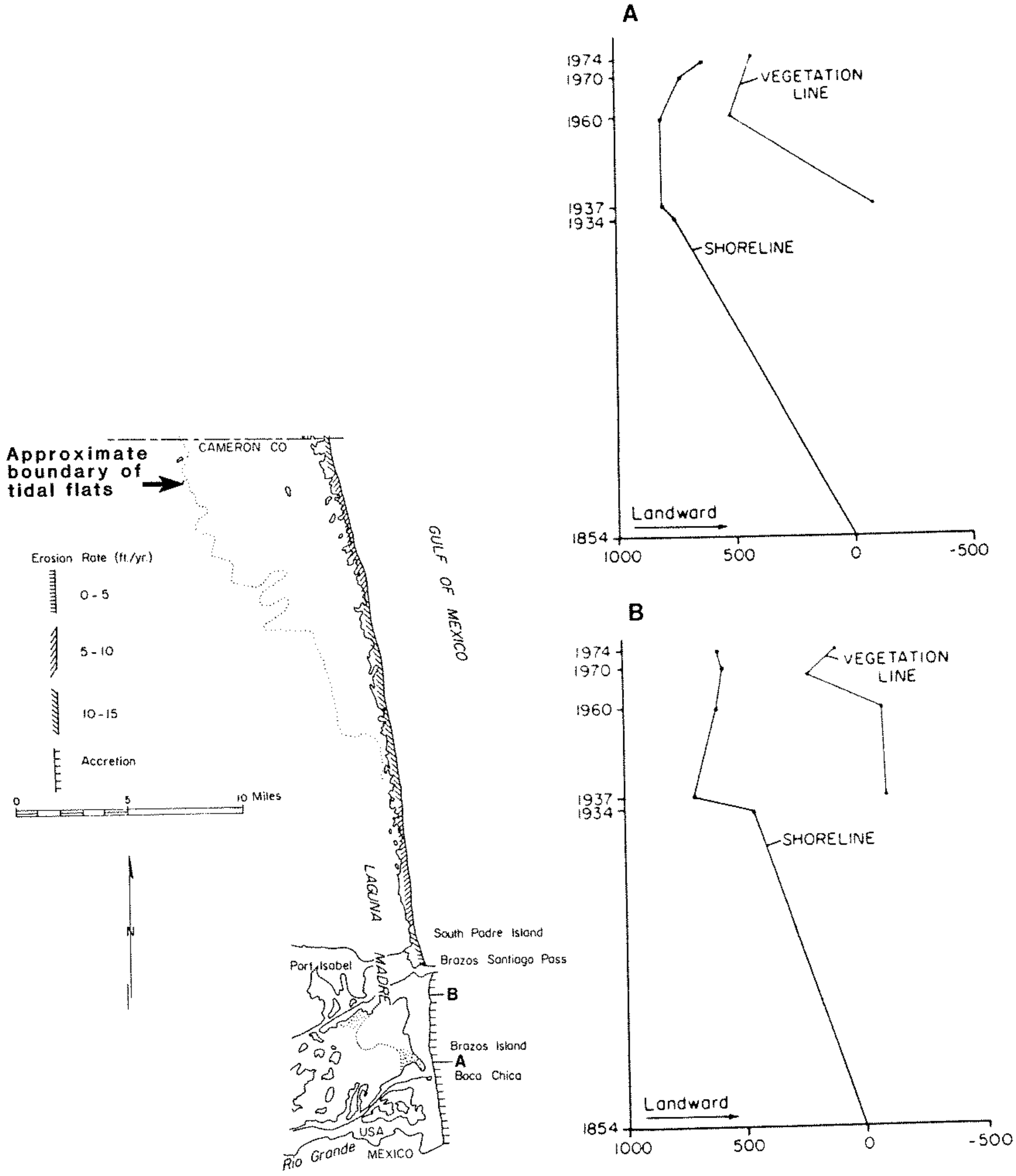

Figure 55. Net historical shoreline changes along the south Texas coast and relative changes in the position of the shoreline and vegetation line at two selected sites, A and B (from Morton and Pieper 1975).

1984, can reduce vegetation density, leading to erosion and aeolian transport of sediment.

The wind regime in South Texas differs depending on season. Throughout much of the year southeast breezes blow onshore, moving sand off the beaches. Winter cold fronts produce winds primarily from the southeast and northeast quadrants. The winds set up strong northward-flowing longshore currents, which also transport sand northward, contributing to the coastal erosion problem in the southern portions of the CBRS unit but, at the same time, accelerating accretion at the northern (jettied) portion of the CBRS unit.

Hurricanes and tropical storms are the single most intense agents of coastal change. The most recent direct storm impact on the Boca Chica area was by Hurricane Allen in 1980 (Suter et al. 1982), but other storms occurred in 1967, 1961, 1945, 1933, 1919, and 1916 (Morton and Paine 1984). The primary effects of the storms are caused by the storm surge and attendant wave scour, and include dune carving, beach erosion, island breaching, washover deposition, and changes in the position of the mouth of the Rio Grande. Many of the hurricanes noted above reactivated the storm channel at Boca Chica (Morton and Pieper 1975).

Sands which built Brazos Island and the Boca Chica area were derived from the erosion of the Rio Grande Delta lobes and were originally brought to the gulf by the river (Brown et al. 1980). Net accretion in this area from 1854 to 1974 (Morton and Pieper 1975; Figure 55) indicates an adequate sediment supply; however, recent erosional trends suggest depletion of both the shelf sediments and river sources because of the damming of the Rio Grande River. In addition, the jetty construction at Brazos Santiago Pass has altered littoral transport of sediment. Subsidence, either human-induced or natural, does not appear to be a major process contributing to erosion based on tide gauge analysis at nearby Port Isabel (Swanson and Thurlow 1973).

Management Implications

Brazos Island and Boca Chica will probably continue to erode at rates of 6 to 20 ft/yr (Morton and Pieper 1975). Boca Chica Pass will most likely be reopened by future hurricanes. Erratic behavior of the Rio Grande River mouth can be expected to continue.

The area is currently undeveloped. Since the State of Texas has removed the Brazos Island State Recreation Area from the jurisdiction of the State Park Department and returned it to the General Land Office (GLO), private development of the area is now possible. In fact, GLO has issued a provisional lease for possible development. Given the historic evidence (e.g., Figure 54), any development has a high probability of experiencing impacts from changes in location of the river mouth.

1982-HISTORIC MAP COMPARISONS

The results of the 1982-historic map comparisons, performed using the FWS geographic information system, are presented in Table 17. Because the photointerpretation methods, the level of interpretation, and the conventions used for drawing the shoreline were not the same for the 1982 maps and the historic maps, the change numbers in this table represent approximate figures that indicate the direction and general magnitude of the change; they should not be considered absolutely accurate. The precise magnitude of the error introduced by the differences in mapping techniques is not known. Visual inspections of overlays of the maps show that the direction of the change (land loss (-), land gain (+)) is accurate, but the magnitude of the change may be inflated, especially for eroding units. In general, the 1982 interpretations include less intertidal land area in their delineations than the older maps (either National Wetland Inventory habitat maps or USGS maps), which biases the change figures towards showing land loss. Eighteen of the 27 CBRS units examined are experiencing substantial erosion within the CBRS boundaries.

It is also important to realize that what is occurring in a CBRS unit is not necessarily what is occurring along the coastal barrier as a whole. This is particularly true for accretionary units. For example, the portion of the coastal barrier included in the Captain Sams Inlet CBRS unit (M08) is the distal end of a barrier spit on the downdrift side of a barrier island. The spit is accreting, but the beaches updrift are gradually eroding.

Conversely, a single net numerical estimate of shoreline changes for an entire large coastal barrier sacrifices knowledge of variations or anomalies within the barrier. For example, Cape San Blas (CBRS Unit P30) is a large unit that has experienced net accretion during the past 40 years (Table 17). That net accretion, however, is the result of very substantial progradation of a small section of the unit's total shoreline (the Cape itself). Most of the length of shoreline is stable or erosional. Obviously, knowledge of shoreline trends on both a large and small scale should be acquired before planning decisions are made.

GENERAL DISCUSSION AND RECOMMENDATONS

The 1982 inventory of CBRS provides a retrievable data base on the entire Coastal Barrier Resources System. The acreage of fastland in each CBRS unit gives a general indication of the developable land in each unit, and the count of structures verifies the undeveloped status of those units. At the same time, however, the classification system used in the inventory was so simplistic that the inventory does not provide much useful habitat data for fish and wildlife management, or sufficiently accurate shoreline data (because of discrepancies in classifying tidal flats). Future inventories should correct these problems.

The 18 case studies and the comparisons of historical maps to 1982 maps show that coastal change is occurring over the entire geographic stretch of CBRS. Most of the change is erosion, but rates and direction of change are highly variable. The finding that most of the shorelines in CBRS are eroding is not surprising. A comprehensive survey of the entire U.S. shoreline (including the Pacific coast and the Great Lakes) reported by May, Dolan, and Hayden (1983) shows a national shoreline erosion rate of 1.3 ft/yr. Along the Atlantic coast, the average erosion rate is 2.6 ft/yr. The Gulf of Mexico coast showed the highest average erosion rate: 5.9 ft/yr.

The general trend of shoreline recession on the Atlantic and gulf coasts is the result of several factors. The damming of many major rivers has significantly reduced the amount of sediment supplied to the coast and available for coastal barrier maintenance. The volume of water in the earth's oceans is increasing because of the melting of the polar glaciers and increased ocean temperatures, and most of the Atlantic and gulf coasts are also slowly subsiding. The combination of these two processes has resulted in about a 1-ft increase in apparent sea level over the last century (Hicks et al. 1983). However, although sea level is rising, it is doing so at a slower rate compared to events in the geologic past (the end of the ice ages). The geological stability of sea level over the past approximately 5,000 years means that onshore transport of sediment from offshore sources has slowed or halted as these sources have either been exhausted or now lie in waters too deep to be reached by wave action.

In the 18 study areas we examined, the construction of coastal stabilization features beyond CBRS boundaries (e.g., jetties and groins) has, in general, resulted in the acceleration of erosion within the CBRS. Stabilization structures alter the dynamic equilibrium of coastal processes. Although barrier systems adjust to the structures in time, sediment that is accumulated around the structures to stabilize one segment of the shoreline must do so at the expense of destabilizing other shoreline segments. The protection of developed areas on coastal barriers (outside of the CBRS) is occurring at the expense of the undeveloped areas that are part of CBRS. Tolerating erosion in CBRS units in order to protect neighboring developed areas may not have been a conscious management decision, but that is what is occurring in some places.

Dredging is a widespread activity along the coast. Numerous artificially created or maintained channels exist, and dredging involves huge volumes of sediment. Dredging affects the littoral zone in two ways. First, it creates sediment sinks (channels) which interrupt the littoral drift system; and second, it removes sediment from the littoral zone and places it in areas that are not available for shoreline maintenance (i.e., subaerial spoil piles). Dredging also changes the characteristics of the submarine sediment surface which alters benthic communities and affects bottom-feeding fishes. The turbidity that dredging causes affects pelagic communities.

Dredging can sometimes be beneficial. For example, sediment can be dredged outside the littoral zone and deposited along the shoreline. Such beach nourishment is expensive but has been successful in slowing erosion rates along some beaches. Also, some spoil islands have become important bird nesting habitats, particularly where traditional nesting areas have been lost. Unpolluted, sand-size spoil should be regarded as a resource, not as a disposal problem. Planned placement of this spoil could help alleviate erosion rather than encourage it.

The widespread impacts of human activities on coastal barriers are clearly shown in the case studies. Only one out of 19 units has not experienced major culturally related impacts (Table 18). Seventeen units have experienced dredging. The vast majority of these units are erosional. Of the two units that are accretional, the dredging activity is minor compared to other human impacts that have resulted in accretion (i.e., downdrift jetties). Fifteen of the 19 units have shoreline stabilization structures in or near the units that impact the shoreline in the CBRS units. Twelve of these 15 units are erosional. Of the three that are accretional, two have structures downdrift of the unit that have encouraged accretion in the unit at the expense of the downdrift coast. The sediment supply to at least 8 of 19 units has been reduced by construction of dams on nearby rivers. All eight of those units are erosional.

Hurricanes and major winter storms (northeasters) are responsible for extensive changes in coastal barriers. All the case study areas have experienced some hurricane or northeast storm impacts or both. The largest amount of sand transport on coastal barriers occurs during these extreme events. Beaches may be eroded 100 ft or more, dunes and marshes may be overwashed, and new inlets may be cut through barriers. The case studies of Dauphin Island, Alabama, Mobile Point, Alabama, and Isles Dernieres, Louisiana, clearly show the major changes that hurricanes can cause. Interspersed between storms, periods of little shoreline change may give a false sense of barrier stability.

While the vast majority of the CBRS units examined here have been undergoing net erosion during at least the last few decades, the loss of habitat due to beach erosion can be compensated by a gain in habitat on the bay side or downdrift end of a barrier if the barrier exists in a natural state. The entire barrier system moves in a landward or longshore direction or both. When barriers change in this manner, net fish and wildlife habitat is not lost, but homes on the beach would be, eventually ending up in the water. Development is the major agent resulting in loss of fish and wildlife habitat.

Table 1. Acreage statistics for Maine CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Lubec Barriers	A01	86.6	27.9	0.0	224.3	0.0	338.8	114.5	0	No	Yes
Baileys Mistake	A01A	29.8	0.0	0.0	0.0	0.0	29.8	29.8	0	No	No
Jasper	A03	5.6	15.7	0.0	38.5	0.0	59.8	21.3	0	Yes	No
Starboard	A03B	11.9	0.7	0.0	14.5	0.1	27.2	12.7	1	No	Yes
Popplestone Beach/Roque Island	A03C	34.9	0.0	0.0	0.0	0.0	34.9	34.9	0	No	No
Seven Hundred Acre Island	A05A	101.8	0.0	0.0	0.0	0.3	102.1	102.1	1	Yes	Yes
Head Beach	A05B	53.7	19.1	0.0	51.8	1.0	125.6	73.8	4	No	Yes
Jenks Landing/Waldo Point	A05C	48.2	0.0	0.0	0.0	0.2	48.4	48.4	1	No	Yes
Cape Elizabeth	A06	29.1	11.4	0.0	2.2	0.0	42.7	40.5	0	Yes	Yes
Scarborough Beach	A07	32.2	64.9	0.0	0.0	0.4	97.5	97.5	2	No	Yes
Crescent Surf	A08	30.1	30.6	0.0	8.5	0.0	69.2	60.7	0	No	Yes
Seapoint	A09	19.4	47.6	0.0	2.4	0.0	69.4	67.0	0	No	No
TOTAL		483.3	217.9	0.0	342.2	2.0	1,045.4	703.2	9		

Table 2. Acreage statistics for Massachusetts CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Clark Pond	C00	9.3	1.3	0.0	24.3	0.0	34.9	10.6	0	No	No
Wingaersheek	C01	64.2	81.1	0.0	65.3	0.0	210.6	145.3	0	No	Yes
Good Harbor Beach	C01A	27.4	16.4	0.0	38.4	0.3	82.5	44.1	1	No	No
Brace Cove	C01B	4.5	8.0	26.4	15.6	0.0	54.5	12.5	0	No	Yes
West Head Beach	C01C	4.1	1.7	0.0	22.8	0.0	28.6	5.8	0	No	Yes
North Scituate	C02	5.0	0.0	15.7	0.0	0.0	20.7	5.0	0	No	No
Rivermoor	C03	18.5	119.6	0.0	82.3	0.0	220.4	138.1	0	No	No
Rexhame	C03A	43.2	5.3	0.0	6.7	0.5	55.8	49.0	2	No	Yes
Plymouth Bay	C04	69.0	97.7	0.0	213.4	2.1	382.2	168.8	6	Yes	Yes
Center Hill Complex	C06	36.7	89.9	19.0	0.0	0.0	145.6	126.6	0	Yes	Yes
Scorton	C08	35.6	12.9	2.2	0.0	0.0	50.7	48.5	0	Yes	No
Sandy Neck	C09	349.8	1,058.6	6.0	1,190.8	2.0	2,607.2	1,410.4	10	Yes	Yes
Freemans Pond	C10	87.3	290.9	6.6	12.9	0.0	397.7	378.2	0	No	Yes
Namskaket Spits	C11	47.0	221.7	0.0	12.6	0.4	281.7	269.1	1	No	No
Boat Meadow	C11A	10.7	65.7	0.0	17.2	0.0	93.6	76.4	0	No	No
Chatham Roads	C12	24.8	77.0	1.2	26.2	0.0	129.2	101.8	0	Yes	Yes
Lewis Bay	C13	168.7	41.9	36.8	424.0	1.7	673.1	212.3	4	Yes	Yes
Squaw Island	C14	14.9	45.3	0.0	17.1	0.0	77.3	60.2	0	Yes	Yes
Centerville	C15	19.3	14.4	0.0	25.7	0.0	59.4	33.7	0	Yes	No
Dead Neck	C16	87.0	14.2	0.0	97.2	0.0	198.4	101.2	0	Yes	No
Popponessett Spit	C17	12.5	3.4	0.0	66.2	0.0	82.1	15.9	0	No	No
Waquoit Bay	C18	376.5	97.6	76.3	587.8	0.0	1,138.2	474.1	0	Yes	Yes
Falmouth Ponds	C18A	11.9	0.0	0.0	13.7	0.2	25.8	12.1	1	Yes	Yes
Black Beach	C19	32.6	124.8	5.1	55.4	0.0	217.9	157.4	0	Yes	Yes
Buzzards Bay Complex	C19A	55.7	146.2	25.5	252.8	0.0	480.2	201.9	0	Yes	Yes
Coatue	C20	233.7	32.9	0.0	283.5	2.1	552.2	268.7	7	No	Yes
Sesachacha Pond	C21	20.2	0.0	0.0	32.5	0.0	52.7	20.2	0	No	Yes
Cisco Beach	C22	14.8	0.0	11.7	0.0	0.0	26.5	14.8	0	No	No
Esther Island	C23	189.6	57.3	0.0	1,272.4	1.4	1,520.7	248.3	6	No	No
Tuckernuck Island	C24	85.5	8.9	0.0	290.1	0.0	384.5	94.4	0	No	Yes
Muskaget Island	C25	196.4	86.8	0.0	2,670.9	0.8	2,954.9	284.0	2	Yes	Yes
Eel Pond Beach	C26	23.5	6.1	0.0	72.1	0.5	102.2	30.1	2	Yes	Yes
Cape Poge	C27	233.5	21.3	0.0	249.0	0.2	504.0	255.0	1	Yes	Yes
South Beach	C28	341.6	44.3	117.7	107.2	0.0	610.8	385.9	0	No	Yes

(continued)

Table 2. Concluded.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Squibnocket Complex	C29	303.4	126.5	107.4	131.5	0.0	668.8	429.9	0	No	Yes
James Pond	C29A	34.9	7.7	0.4	43.6	0.0	86.6	42.6	0	Yes	No
Mink Meadows	C29B	24.6	3.2	8.5	5.6	0.0	41.9	27.8	0	Yes	No
Elizabeth Islands	C31	156.6	39.7	64.7	343.9	0.7	605.6	197.0	2	Yes	No
West Sconticut Neck	C31A	55.6	62.7	1.2	159.2	0.5	279.2	118.8	2	No	No
Harbor View	C31B	10.1	33.7	0.0	25.5	0.0	69.3	43.8	0	No	Yes
Misham Point	C32	9.6	14.6	0.0	97.8	0.0	122.0	24.2	0	Yes	Yes
Little Beach	C33	79.8	112.4	61.5	24.4	4.0	282.1	196.2	23	No	Yes
Horseneck Beach	C34	143.6	39.7	2.6	164.1	10.7	360.7	194.0	64	No	Yes
Cedar Cove	C34A	6.3	2.9	0.0	7.3	0.1	16.6	9.3	1	No	No
TOTAL		3,779.5	3,336.3	596.5	9,249.0	28.2	16,989.5	7,144.0	135		

Table 3. Acreage statistics for Rhode Island CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Little Compton Ponds	D01	344.9	81.5	227.5	700.7	3.8	1,358.4	430.2	10	Yes	Yes
Fogland Marsh	D02	44.5	73.7	16.0	57.6	2.7	194.5	120.9	7	Yes	Yes
Prudence Island Complex	D02B	52.8	143.5	0.0	104.0	0.2	300.5	196.5	2	Yes	Yes
West Narragansett Bay Complex	D02C	9.3	8.0	0.0	3.8	0.0	21.1	17.3	0	No	Yes
Card Ponds	D03	63.2	10.4	0.0	52.1	2.5	128.2	76.1	10	No	No
Green Hill Beach	D04	62.9	0.0	0.0	185.8	1.5	250.2	64.4	4	No	No
East Beach	D05	73.4	124.5	1.1	202.8	0.6	402.4	198.5	2	Yes	Yes
Quonochontaug Beach	D06	121.7	100.4	0.5	191.6	0.0	414.2	222.1	0	No	Yes
Maschaug Ponds	D07	48.0	0.0	44.9	0.0	0.0	92.9	48.0	0	No	No
Napatree	D08	120.1	0.0	0.0	1,610.4	0.0	1,730.5	120.1	0	No	No
Block Island	D09	168.6	55.8	96.2	112.8	1.5	434.9	225.9	4	No	Yes
TOTAL		1,109.4	597.8	386.2	3,221.6	12.8	5,327.8	1,720.0	39		

Table 4. Acreage statistics for Connecticut CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks, or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Wilcox Beach	E01	6.7	11.4	0.0	6.3	0.0	24.4	18.1	0	No	Yes
Ram Island	E01A	15.9	0.0	0.0	237.8	1.2	254.9	17.1	5	Yes	Yes
Goshen Cove	E02	18.2	20.8	0.0	23.9	0.0	62.9	39.0	0	No	Yes
Jordon Cove	E03	11.6	0.0	0.0	23.1	0.0	34.7	11.6	0	No	No
Niantic Bay	E03A	6.9	11.3	2.0	23.1	0.0	20.2	18.2	0	No	No
Lynde Point	E03B	58.6	9.5	0.0	98.5	4.2	170.8	72.3	18	Yes	Yes
Menunketesuck Island	E04	4.5	7.1	0.0	148.8	0.0	160.4	11.6	0	No	No
Hammonasset Point	E05	8.8	1.6	0.0	51.3	0.1	61.8	10.5	1	No	No
Milford Point	E07	15.2	21.6	0.0	15.4	0.0	52.2	36.8	0	Yes	No
Fayerweather Island	E08A	13.4	0.0	0.0	227.5	0.2	241.1	13.6	1	Yes	No
Norwalk Islands	E09	165.6	11.1	0.0	1,247.6	1.5	1,425.8	178.2	8	Yes	No
TOTAL		325.4	94.4	2.0	2,080.2	7.2	2,509.2	427.0	33		

Table 5. Acreage statistics for New York CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Fishers Island Barriers	F01	16.6	10.2	0.0	14.5	0.0	41.3	26.8	0	No	No
Eatons Neck	F02	15.8	0.0	0.0	21.1	0.0	36.9	15.8	0	No	No
Crane Neck	F04	29.2	100.3	18.0	0.0	0.0	147.5	129.5	0	Yes	No
Old Field Beach	F05	148.6	28.4	0.0	232.8	0.0	409.8	177.0	0	Yes	Yes
Shelter Island Barriers	F06	81.8	39.6	0.0	95.2	1.2	217.8	122.6	3	Yes	Yes
Sammys Beach	F08A	96.0	60.9	0.3	193.8	2.3	353.3	159.2	14	Yes	Yes
Acabonack Harbor	F08B	53.6	17.9	0.0	100.7	0.0	172.2	71.5	0	Yes	Yes
Gardiners Island Barriers	F09	165.9	148.3	202.3	1,083.6	0.0	1,600.1	314.2	0	Yes	No
Napeague	F10	145.8	28.4	0.0	34.2	5.8	214.2	180.0	7	No	Yes
Mecox	F11	41.4	8.7	49.1	0.0	0.6	99.8	50.7	1	No	Yes
Southampton	F12	186.7	25.3	0.0	333.6	3.1	548.7	215.1	5	No	Yes
Tiana Beach	F13	128.9	554.7	0.0	102.2	7.3	793.1	690.9	24	No	Yes
TOTAL		1,110.3	1,022.7	269.7	2,211.7	20.3	4,634.7	2,153.3	54		

Table 6. Acreage statistics for Delaware CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Broadkill Beach Complex	H00	362.5	952.3	18.5	25.8	12.4	1,371.5	1,327.2	59	Yes	Yes
North Bethany Beach	H01	141.3	40.7	0.0	10.1	1.1	193.2	183.1	2	Yes	Yes
TOTAL		503.8	993.0	18.5	35.9	13.5	1,564.7	1,510.3	61		

Table 7. Acreage statistics for Virginia CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Assawoman Island	K01	281.9	523.4	12.3	198.3	0.0	1,015.9	805.3	0	Yes	No
Cedar Island	K03	612.7	4,626.8	0.0	4,019.1	22.0	9,280.6	5,261.5	24	Yes	Yes
Little Cobb Island	K04	2.1	3.7	0.0	378.1	0.0	383.9	5.8	0	Yes	No
Fishermans Island	K05	228.9	256.3	3.6	128.0	0.5	617.3	485.7	1	No	Yes
TOTAL		1,125.6	5,410.2	15.9	4,723.5	22.5	11,297.7	6,558.3	25		

Table 8. Acreage statistics for North Carolina CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Currituck Banks	L01	1,514.1	2,263.1	105.8	3,372.1	1,005.0	8,260.1	4,782.2	239	Yes	Yes
Hatteras Island	L03	134.8	81.6	0.0	112.3	0.0	328.7	216.4	0	No	Yes
Shackleford Banks	L03A	2,009.0	668.1	1.8	5,816.2	32.5	8,527.6	2,709.6	58	Yes	No
Onslow Beach Complex	L05	1,247.9	1,226.2	0.0	485.7	2.2	2,962.0	2,476.3	8	No	Yes
Topsail	L06	894.6	1,127.1	79.6	509.0	7.5	2,617.8	2,029.2	25	Yes	Yes
Lea Island Complex	L07	407.7	2,280.8	0.0	677.8	7.0	3,373.3	2,695.5	8	No	Yes
Wrightsville Beach	L08	117.4	141.0	0.0	48.9	0.0	307.3	258.4	0	No	Yes
Masonboro Island	L09	1,097.4	2,416.0	10.7	1,677.1	0.8	5,202.0	3,514.2	3	Yes	Yes
TOTAL		7,422.9	10,203.9	197.9	12,699.1	1,055.0	31,578.8	18,681.8	341		

Table 9. Acreage statistics for South Carolina CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Waites Island Complex	M01	527.8	929.2	0.0	958.1	0.8	2,415.9	1,457.8	2	Yes	Yes
Litchfield Beach	M02	26.7	2.4	0.0	42.7	0.0	71.8	29.1	0	No	No
Pawleys Inlet	M03	45.4	58.8	0.0	26.1	0.0	130.3	104.2	0	Yes	Yes
Debidue Beach	M04	232.3	208.0	0.0	166.7	0.3	607.3	440.6	1	Yes	Yes
Dewees Island	M05	101.7	698.8	15.4	166.6	0.0	982.5	800.5	0	Yes	Yes
Morris Island Complex	M06	75.0	2,144.2	37.0	439.9	0.0	2,696.1	2,219.2	0	No	No
Bird Key Complex	M07	376.2	392.1	0.0	890.3	0.0	1,658.6	768.3	0	No	Yes
Captain Sams Inlet	M08	155.8	162.2	0.0	174.5	0.0	492.5	318.0	0	No	No
Edisto Complex	M09	192.1	1,044.3	0.0	159.2	0.0	1,395.6	1,236.4	0	No	Yes
Otter Island	M10	303.2	1,694.2	0.0	728.4	0.0	2,725.8	1,997.4	0	No	No
Harbor Island	M11	36.0	174.7	16.9	22.4	0.0	250.0	210.7	0	No	Yes
St. Phillips Island	M12	1,400.6	7,439.7	0.0	2,589.1	7.7	11,437.1	8,848.0	10	Yes	Yes
Daufuskie Island	M13	1,029.3	998.0	0.0	328.3	0.0	2,355.6	2,027.3	0	Yes	No
TOTAL		4,502.1	15,946.6	69.3	6,692.3	8.8	27,219.1	20,457.5	13		

Table 10. Acreage statistics for Georgia CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Little Tybee Island	N01	905.3	5,275.1	65.5	1,642.7	0.0	7,888.6	6,180.4	0	No	No
Wassaw Island	N01A	146.6	132.0	0.0	24.4	11.2	314.2	289.8	10	No	Yes
Little St. Simons Island	N03	1,848.0	8,494.3	0.0	2,264.0	3.7	12,610.0	10,346.0	3	No	Yes
Sea Island	N04	266.2	46.0	0.0	126.0	0.0	438.2	312.2	0	No	Yes
Little Cumberland Island	N05	270.5	827.3	0.0	539.9	2.0	1,639.7	1,099.8	3	No	Yes
Cumberland Island	N06	1,669.5	4,214.5	0.0	4,294.8	3.4	10,182.2	5,887.4	4	Yes	Yes
TOTAL		5,106.1	18,989.2	65.5	8,891.8	20.3	33,072.9	24,115.6	20		

Table 11. Acreage statistics for Florida CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Talbot Islands Complex	P02	1,912.4	2,681.0	0.0	2,200.2	0.7	6,794.3	4,594.1	1	Yes	Yes
Usinas Beach	P04A	58.9	100.1	0.0	111.6	8.8	279.4	167.8	15	No	Yes
Conch Island	P05	743.9	86.7	0.0	356.5	0.0	1,187.1	830.6	0	Yes	Yes
Matanzas River	P05A	53.0	59.4	2.7	48.6	1.6	165.3	114.0	8	Yes	Yes
Ormond-by-the-Sea	P07	575.1	89.4	0.0	70.9	0.0	735.4	664.5	0	Yes	Yes
Ponce Inlet	P08	360.3	64.4	0.0	443.1	3.6	871.4	428.3	10	Yes	Yes
Coconut Point	P09A	252.7	0.0	2.0	509.4	5.4	769.5	258.1	24	Yes	Yes
Vero Beach	P10	90.9	17.9	1.5	215.8	11.1	337.2	119.9	36	Yes	Yes
Blue Hole	P10A	537.4	1,059.8	102.8	1,358.2	16.2	3,074.4	1,613.4	31	No	Yes
Hutchinson Island	P11	419.6	2,113.8	198.2	3,102.4	2.1	5,836.1	2,535.5	10	No	Yes
Hobe Sound	P12	25.6	0.0	0.0	0.0	0.0	25.6	25.6	0	No	No
North Beach	P14A	75.8	0.0	52.3	0.0	0.0	128.1	75.8	0	Yes	Yes
Cape Romano	P15	448.6	1,858.4	47.5	1,106.2	0.9	3,461.6	2,307.9	2	Yes	No
Keewaydin Island	P16	1,071.1	878.2	137.5	846.1	13.6	2,946.5	1,962.9	36	Yes	Yes
Lovers Key Complex	P17	182.1	320.9	0.5	692.3	0.3	1,196.1	503.3	1	No	Yes
Bodwitch Point	P17A	14.2	0.0	0.0	55.6	0.3	70.1	14.5	1	No	Yes
Sanibel Island	P18	84.4	117.2	0.0	223.9	2.4	427.9	204.0	3	No	Yes
North Captiva Island	P19	64.3	53.4	0.0	212.2	0.0	329.9	117.7	0	No	Yes
Cayo Costa	P20	834.6	343.6	11.6	1,238.1	13.6	2,441.5	1,191.8	18	Yes	Yes
Bocilla Island	P21	507.8	205.8	16.6	828.5	7.0	1,565.7	720.6	25	Yes	Yes
Manasota Island	P21A	24.5	36.9	0.2	8.5	0.3	70.4	61.7	1	Yes	Yes
Casey Key	P22	88.7	52.8	1.6	253.6	0.6	397.3	142.1	2	No	Yes
Longboat Key	P23	47.5	6.4	0.0	180.5	0.4	234.8	54.3	2	Yes	Yes
The Reefs	P24	91.7	107.2	0.0	1,217.8	0.3	1,417.0	199.2	1	Yes	No
Mandalay Point	P24A	40.4	0.0	1.2	18.3	0.0	59.9	40.4	0	No	No
Atsena Otie Key	P25	57.1	43.5	0.3	650.9	0.0	751.8	100.6	0	No	No
Pepperfish Keys	P26	55.7	81.3	0.0	567.1	0.3	704.4	137.3	1	No	No
Ochlockonee Complex	P27A	179.6	245.6	6.9	121.7	0.0	553.8	425.2	0	No	Yes
Dog Island	P28	348.9	259.1	14.6	905.3	45.8	1,573.7	653.8	55	Yes	Yes
Cape San Blas	P30	2,485.9	478.6	5.2	1,796.2	37.8	4,803.7	3,002.3	93	Yes	Yes
St. Andrew Complex	P31	3,639.1	689.1	75.1	7,694.4	23.8	12,121.5	4,352.0	25	Yes	Yes
Four Mile Village	P31A	1,129.3	521.2	189.8	32.1	6.6	1,879.0	1,657.1	8	No	Yes
Moreno Point	P32	2,695.1	426.3	25.5	1,190.0	27.6	4,364.5	3,149.0	39	No	Yes
TOTAL		19,196.2	12,998.0	893.6	28,256.0	231.1	61,574.9	32,425.3	448		

Table 12. Acreage statistics for Alabama CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Mobile Point	Q01	1,895.4	1,868.2	0.0	2,595.3	9.8	6,368.7	3,773.4	16	Yes	Yes
Pelican Island	Q01A	121.3	0.0	0.0	1,376.8	0.0	1,498.1	121.3	0	No	No
Dauphin Island	Q02	913.9	182.1	3.7	1,711.5	0.0	2,811.2	1,096.0	0	Yes	Yes
TOTAL		2,930.6	2,050.3	3.7	5,683.6	9.8	10,678.0	4,990.7	16		

Table 13. Acreage statistics for Mississippi CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Round Island	R01	50.4	16.8	0.0	876.5	0.0	943.7	67.2	0	No	No
Belle Fontaine Point	R01A	47.4	538.0	0.5	94.6	1.4	681.9	586.8	8	No	No
Deer Island	R02	238.9	263.8	0.0	1,510.7	2.2	2,015.6	504.9	6	Yes	Yes
Cat Island	R03	216.6	332.5	0.0	119.1	0.0	668.2	549.1	0	No	Yes
TOTAL		553.3	1,151.1	0.5	2,600.9	3.6	4,309.4	1,708.0	14		

Table 14. Acreage statistics for Louisiana CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Bastian Bay Complex	S01	205.3	537.7	9.5	908.0	0.4	1,660.9	743.4	1	Yes	No
Bay Joe Wise Complex	S01A	259.6	312.1	56.4	452.6	0.3	1,081.0	572.0	1	No	No
Grande Terre Islands	S02	106.1	193.5	0.0	739.7	0.8	1,040.1	300.4	1	No	No
Caminada	S03	119.3	126.7	6.2	473.7	0.2	726.1	246.2	1	No	Yes
Bay Champagne	S04	273.9	781.0	36.1	1,461.1	1.3	2,553.4	1,056.2	4	Yes	Yes
Timbalier Island	S05	632.8	1,066.8	30.4	6,421.7	0.0	8,151.7	1,699.6	0	No	No
Isles Dernieres	S06	553.4	1,813.4	47.7	11,344.9	13.7	13,773.1	2,380.5	19	No	No
Point au Fer	S07	226.4	10,548.2	527.3	2,707.1	1.6	14,010.6	10,776.2	2	No	No
Chenier au Tigre	S08	52.9	557.9	0.0	6.2	3.9	620.9	614.7	4	No	No
Rollover	S09	22.3	99.8	0.0	0.0	0.0	122.1	122.1	0	No	No
Mermentau River	S10	491.0	7,179.9	164.3	1,936.8	1.8	9,773.8	7,672.7	3	Yes	No
Sabine	S11	1,513.8	4,042.3	136.4	0.0	37.1	5,729.6	5,593.2	23	Yes	No
TOTAL		4,456.8	27,259.3	1,014.3	26,451.8	61.1	59,243.3	31,777.2	59		

Table 15. Acreage statistics for Texas CBRS units. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures; JDG=Jetty/Dock/Groin. Only presence or absence of jetties, docks or groins, and roads is indicated.

Unit Name	ID Code	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)	JDG	Roads
Sea Rim	T01	2,236.1	12,212.0	869.4	264.6	90.2	15,672.3	14,538.3	72	Yes	Yes
High Island	T02A	954.3	20,413.8	761.1	97.3	14.7	22,241.2	21,382.8	11	Yes	Yes
Bolivar Peninsula	T03A	817.6	5,765.7	9.1	265.4	84.1	6,941.9	6,667.4	97	No	Yes
Follets Island	T04	872.9	2,122.1	64.1	123.7	9.6	3,192.4	3,004.6	23	Yes	Yes
Brazos River Complex	T05	1,342.9	2,195.9	494.7	75.3	1.0	4,109.8	3,539.8	1	No	Yes
Sargent Beach	T06	298.2	308.0	6.8	3.8	0.0	616.8	606.2	0	No	No
Matagorda Peninsula	T07	10,385.1	14,000.8	675.5	528.0	24.1	25,613.5	24,410.0	22	Yes	Yes
San Jose Island Complex	T08	19,102.9	26,525.6	1,799.4	1,881.7	49.8	49,359.4	45,678.3	48	Yes	Yes
North Padre Island	T10	3,717.2	1,389.8	0.0	21.8	3.5	5,132.3	5,110.5	2	No	Yes
South Padre Island	T11	5,722.0	40,130.4	271.7	36.0	2.8	46,162.9	45,855.2	2	No	Yes
Boca Chica	T12	1,243.6	1,251.7	17.3	0.0	9.8	2,522.4	2,505.1	6	No	Yes
TOTAL		46,692.8	126,315.8	4,969.1	3,297.6	289.6	181,564.9	173,298.2	276		

Table 16. Total CBRS acreage by State. Key: FL=Fastland; WL=Wetland; IOW=Interior Open Water; OW=Open Water; D=Developed; Str=Structures.

State	FL (acres)	WL (acres)	IOW (acres)	OW (acres)	D (acres)	Total (acres)	Total w/o water (acres)	Str (no.)
Maine	483.3	217.9	0.0	342.2	2.0	1,045.4	703.2	9
Massachusetts	3,779.5	3,336.3	596.5	9,249.0	28.2	16,989.5	7,144.0	135
Connecticut	325.4	94.4	2.0	2,080.2	7.2	2,509.2	427.0	33
Rhode Island	1,109.4	597.8	386.2	3,221.6	12.8	5,327.8	1,720.0	39
New York	1,110.3	1,022.7	269.7	2,211.7	20.3	4,634.7	2,153.3	54
Delaware	503.8	993.0	18.5	35.9	13.5	1,564.7	1,510.3	61
Virginia	1,125.6	5,410.2	15.9	4,723.5	22.5	11,297.7	6,558.3	25
North Carolina	7,422.9	10,203.9	197.9	12,699.1	1,055.0	31,578.8	18,681.8	719
South Carolina	4,502.1	15,946.6	69.3	6,692.3	8.8	27,219.1	20,457.5	13
Georgia	5,106.1	18,989.2	65.5	8,891.8	20.3	33,072.9	24,115.6	20
Florida	19,196.2	12,998.0	893.6	28,256.0	231.1	61,574.9	32,425.3	448
Alabama	2,930.6	2,050.3	3.7	5,683.6	9.8	10,678.0	4,990.7	16
Mississippi	553.3	1,151.1	0.5	2,600.9	3.6	4,309.4	1,708.0	44
Louisiana	4,456.8	27,259.3	1,014.3	26,451.8	61.1	59,243.3	31,777.2	59
Texas	46,692.8	126,315.8	4,969.1	3,297.6	289.6	181,564.9	173,298.2	276
TOTAL	99,298.1	226,586.5	8,502.7	116,437.2	1,785.8	452,610.3	327,670.4	1,951

Table 17. Areal changes (in acres) in selected CBRS units, ca. 1940's to 1982.

Unit Name	ID Code	Land Area[a] 1982	Land Area Pre-1982[b]	Land Area Change	Percent Land Area Change	Annual Percent Land Area Change
Maine						
Jasper	A03	21.42	10.92 (1946)	+10.49	+96.0	+2.67
Massachusetts						
Plymouth Bay	C04	168.44	129.25 (1947, 1950)	+39.19	+30.32	+0.91
Popponesset Spit	C17	16.18	20.93 (1949)	-4.75	-22.6	-0.68
Rhode Island						
Quonochontaug Beach	D06	225.11	225.79 (1953)	0	0 (within error margin)	0
Maschaug Ponds	D07	48.93	52.84 (1953)	-3.91	-7.39	-0.25
Connecticut						
Jordan Cove	E03	11.60	19.74 (1937)	-8.14	-41.2	-0.86
Fayerweather Island	E08	13.27	14.50 (1949)	-1.23	-8.48	-0.26
Delaware						
North Bethany Beach	H01	183.44	201.85 (1943)	-18.01	-8.92	-0.23
Virginia						
Little Cobb Island	K04	6.33	77.17 (1942)	-70.84	-91.79	-2.29

(continued)

Table 17. continued.

Unit Name	ID Code	Land Area[a] 1982	Land Area Pre-1982[b]	Land Area Change	Percent Land Area Change	Annual Percent Land Area Change
North Carolina						
Hatteras Island	L03	215.79	217.32 (1946)	0	0 (within error margin)	0
Shackleford Banks	L03A	2,709.69	2,377.67 (1946)	+332.02	+13.96	+0.39
Topsail	L06	2,029.39	2,026.26 (1949,1952)	0	0 (within error margin)	0
South Carolina						
Captain Sams Inlet	M08	322.16	287.09 (1959)	+35.07	+12.21	+0.53
Edisto Complex	M09	1,233.77	1,324.14 (1959)	-90.37	- 6.82	-0.30
Georgia						
Cumberland Island	N06	5,856.11	5,596.81 (1958)	+259.30	+ 4.63	+0.19
Florida						
Ormond-by-the Sea	P07	663.32	747.97 (1956)	-84.65	-11.31	-0.43
Coconut Point	P09A	255.26	281.24 (1947)	-25.98	- 9.23	-0.26
Blue Hole	P10A	1,602.75	1,697.29 (1946)	-94.54	- 5.57	-0.16

(continued)

Table 17. continued.

Unit Name	ID Code	Land Area[a] 1982	Land Area Pre-1982[b]	Land Area Change	Percent Land Area Change	Annual Percent Land Area Change
Hutchinson Island	P11	2,047.19	2,142.68 (1946)	-95.49	- 4.45	-0.12
Cape San Blas	P30	3,002.43	2,915.70 (1943)	+86.73	+ 2.97	+0.08
Moreno Point	P32	3,147.90	3,267.81 (1955)	-119.91	- 3.66	-0.13
Alabama						
Mobile Point	Q01	3,769.79[c]	3,945.49[d] (1979)	-175.70	- 4.45	-1.48
		3,945.49 (1979)	3,969.65 (1956)	0	0 (within error margin)	0
Mississippi						
Deer Island	R02	505.18	544.05 (1952)	- 38.86	- 7.14	-0.23
Cat Island	R03	549.04	804.77 (1956)	-255.73	-31.77	-1.22
Louisiana						
Isles Dernieres	S06	2,381.55	4,616.67 (1956)	-2,235.12	-48.41	-1.86
Texas						
Bolivar Peninsula	T03A	6,667.19	6,759.65 (1956)	-92.46	- 1.37	-0.05
Follets Island	T04	3,005.59[e]	2,914.18 (1956)	+91.41	+ 3.13	+0.12

(continued)

Table 17. Concluded.

Unit Name	ID Code	Land Area[a] 1982	Land Area Pre-1982[b]	Land Area Change	Percent Land Area Change	Annual Percent Land Area Change
Boca Chica	T12	2,519.6 (1948,1950)	2,580.97	-61.32	- 2.37	-0.07

[a]Land area includes the following habitats: fastlands, wetlands, and developed land. Excluded habitats include: open water and inland open water. These numbers differ slightly from the total without water reported in earlier tables because different computational procedures were used.

[b]The number(s) in parentheses indicate the year of the earlier map base. Multiple dates indicate more than one quandrangle map was required to cover the entire CBRS unit and these different maps had different data dates.

[c]After Hurricane Frederick.

[d]Before Hurricane Frederick.

[e]Does not include extensive erosion caused by Hurricane Alicia (1983).

Table 18. Summary of human perturbations of case study CBRS units.

Human Perturbation	No. of units by condition Eroding	 Accreting	 Stable	Total
Dredging	15	2	0	17
Structures-updrift	7	1	0	8
Structures-downdrift	5	2	0	7
Structures-within	7	2	0	9
Dams	8	0	0	8
None	1	0	0	1
Number of case study units per condition	16	3	0	19

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