

*HISTORICAL DEVELOPMENT OF OCEAN CITY INLET EBB SHOAL
AND ITS EFFECT ON NORTHERN ASSATEAGUE ISLAND*



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Preface

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1 Introduction

Assateague Island extends from Ocean City Inlet, Maryland southward to Chincoteague Inlet, Virginia. The northern end of the island is managed by the National Park Service (NPS) as Assateague Island National Seashore. Dynamic physical and ecological changes to northern Assateague Island, associated with inlet stabilization in 1933, have prompted the NPS to investigate a variety of management alternatives to mitigate human-induced impacts and to restore natural geomorphological processes to this area. A workshop was convened in February, 1986 by the NPS to assemble experts to examine erosion control management alternatives. Workshop participants concluded that unless action was taken, profound changes would continue to take place, eventually leading to the demise of northern Assateague Island and alteration of Ocean City Inlet and the shoreline of Sinepuxent Bay. Recent evidence, nonetheless, suggests the rate of physical change along northern Assateague Island may be slowing, and management alternative scenarios previously considered at the workshop may need modification. However, insufficient data are available on which a detailed assessment of present physical and ecological changes can be based. Consequently, the NPS has initiated a monitoring program for the northern portion of Assateague Island. A key objective of this monitoring program is to investigate the relationship between Ocean City Inlet ebb shoal growth and historical shoreline changes along northern Assateague Island.

Prior to 1933, Assateague Spit, in the vicinity of what is now northern Assateague Island, was migrating landward at approximately 2 m/yr. On August 23, 1933, hurricane forces caused a breach in Fenwick Island, creating Ocean City Inlet (Truitt 1967). Ocean City Inlet separated Fenwick Island to the north and Assateague Island to the south. Inlet stabilization began one month later, with jetty construction completed in 1935. The jetties and the new inlet interrupted longshore sediment transport patterns. The result was sand accumulation on Ocean City beaches north of the north jetty, development of an offshore ebb-tidal shoal and a flood tidal shoal in the adjacent bay, and severe downdrift erosion along northern Assateague Island. As the northern jetty and ebb-shoal trapped littoral materials, shoreline recession rates along northern Assateague Island exceeded 12.2 m/yr (Leatherman 1979). Although the fillet at the north jetty filled, the ebb-shoal also acted as a sink for sediment perpetuating the high rate of shoreline recession to the south. During this time, the shoreline and dunes of northern Assateague migrated landward until overwash penetrated across the full width of the barrier. Since that time, decreasing barrier width, lowering of dunes, and frequent overwash has altered ecological characteristics as well. Prior to 1986, investigations of erosional rates for northern Assateague have failed to account for and/or dismissed the occurrence of natural bypassing of sediment around the ebb delta to the island itself. Recent data and observations by Underwood and Anders 1989, suggest that ebb shoal growth may be slowing, possibly due to bypassing of littoral materials to northern Assateague, which would reduce the rate of shoreline recession/island migration (Underwood and Anders 1989, USACE 1986a).

In July 1990, representatives from the NPS visited the U.S. Army Engineer Waterways Experiment Station's Coastal Engineering Research Center (CERC) to discuss physical and ecological changes presently occurring along northern Assateague Island. Ongoing projects conducted by CERC at Ocean City Inlet, at Ocean City, MD, and on the Ocean City Inlet ebb-shoal could directly supplement any data collection program undertaken by the NPS. In 1992, the NPS contracted with the Coastal Structures and Evaluation Branch (CSEB) of the Engineering Development Division (EDD), Coastal Engineering Research Center to continue work presently being conducted on Ocean City Inlet ebb-shoal growth and correlation to historical shoreline erosion trends along northern Assateague Island. If sand is presently bypassing the ebb shoal and being transported downdrift, patterns of growth and deposition should be measurable along the adjacent shoreline of northern Assateague Island. The verification and/or quantification of the Ocean City Inlet ebb shoal bypassing mechanism is critical to predicting the long term future of northern Assateague Island. Examination of temporal and spatial bathymetric changes across the shoal is the first step to evaluation of this "sand bridge" concept. The historical development of Ocean City Inlet ebb shoal (1933-1990), and its relationship to shoreline changes along northern Assateague Island (1849-1989) is the focus of this investigation.

2 Study Objectives

- (1) Provide a comprehensive history of the growth and morphodynamics of the Ocean City Inlet ebb shoal from 1933 to 1990.
- (2) Provide quantitative shoreline change rates and erosion and accretion patterns along northern Assateague Island from 1849 to 1989.
- (3) Define the current impact of the shoal on northern Assateague Island, and predict the future evolution of the shoal and its impact on northern Assateague Island.

Selecting the best management strategy for the northern end of Assateague Island and monitoring its success demands careful, scientifically valid, field data collection and interpretation. Measurements of shoal morphology and shoreline position change will provide a picture of the evolutionary sequences of shoal development and trends in areas of erosion and deposition. An understanding of the local sedimentological processes will provide improved planning and design for physical monitoring programs, provide updated information for a sediment budget, and quantification of the ebb shoal's importance in controlling erosion on adjacent northern Assateague Island. Cumulatively, this detailed examination of sediment and erosion/accretion patterns across the ebb shoal and adjacent shoreline, will provide essential information for development of a long-term physical monitoring program for northern Assateague Island.

3 Regional Setting

Ocean City Inlet is located at approximately 38°20' N and 75°05' W, on the Delaware-Maryland-Virginia (Delmarva) peninsula, approximately 200 km north-northeast of Norfolk, Virginia, and 160 km south-southeast of Baltimore, Maryland (Figure 1). Ocean City Inlet separates Ocean City to the north and Assateague Island to the south. Before the storm of 1933, Ocean City and Assateague Island were adjacent portions of Fenwick Island spit, originating from the Bethany Beach Delaware headland. During this time, Assateague was called Assateague Island, as distinct from Fenwick, although technically neither were islands (USDOI-NPS 1985).

Fenwick Island is approximately 16 km long, averages 1.6 km in width near the Delaware-Maryland border to approximately .40 km wide at Ocean City Inlet. Fenwick Island and the northern part of Assateague Island lie within Worcester County, Maryland. Elevations are generally less than 3 m above mean low water (MLW). Fenwick Island is backed north to south by Assawoman and Isle of Wight bays, respectively. The oceanside shoreline has a smooth curving configuration while the bay shoreline is scalloped and lobate in form. These barrier shoreline characteristics are indicative of past washover events and relict tidal inlet locations (Halsey 1978, McBride 1986). The southern portion of Fenwick's nearshore is triangular in shape, and has accreted seaward a distance of approximately 180 m since inlet stabilization in 1935. A divergent nodal point exists near the Maryland/Delaware state line, reversing net southerly longshore littoral drift to a northerly direction, as evidenced by growth of the fillet adjacent to the south jetty at Indian River Inlet, Delaware. At Ocean City, the net transport is in the opposite direction. Mann and Dalrymple (1986), state that "the nodal point has migrated southward from Bethany Beach to York Beach" (Figure 2). However, they also point to annual variations in nodal point position, "from Indian River Inlet to the north, and as far south as the northern sections of Ocean City, Maryland".

Assateague Island extends from Ocean City Inlet, Maryland southward to Chincoteague Inlet, Virginia. The southern part of Assateague lies in Accomac County, Va. The Maryland portion of the island, and a contiguous envelope of adjacent waters surrounding the entire island, is managed by the NPS as Assateague Island National Seashore. The Virginia portion of the island is under jurisdiction of the United States Fish and Wildlife Service (USFWS) and managed as Chincoteague National Wildlife Refuge. The entire island is approximately 58 km long, averages 4.5 km wide, and is the longest unbroken coastal barrier along the Delmarva Peninsula (Geotlle 1978). Assateague Island is separated from the Eastern Shore mainland by Sinepuxent and Chincoteague Bays. Assateague's oceanside shoreline configuration is straight and curving as is Fenwick Island's, while the inner shoreline is very irregular from washover events and relict tidal deltas. Because this shoreline remains in its natural state, pronounced morphological features are evident from aerial photography such as dunes, washover deltas, reticulate bars, washarounds, and sluiceways. The southern most 8 km (Fishing Point, Va) has been elongating by southwesterly

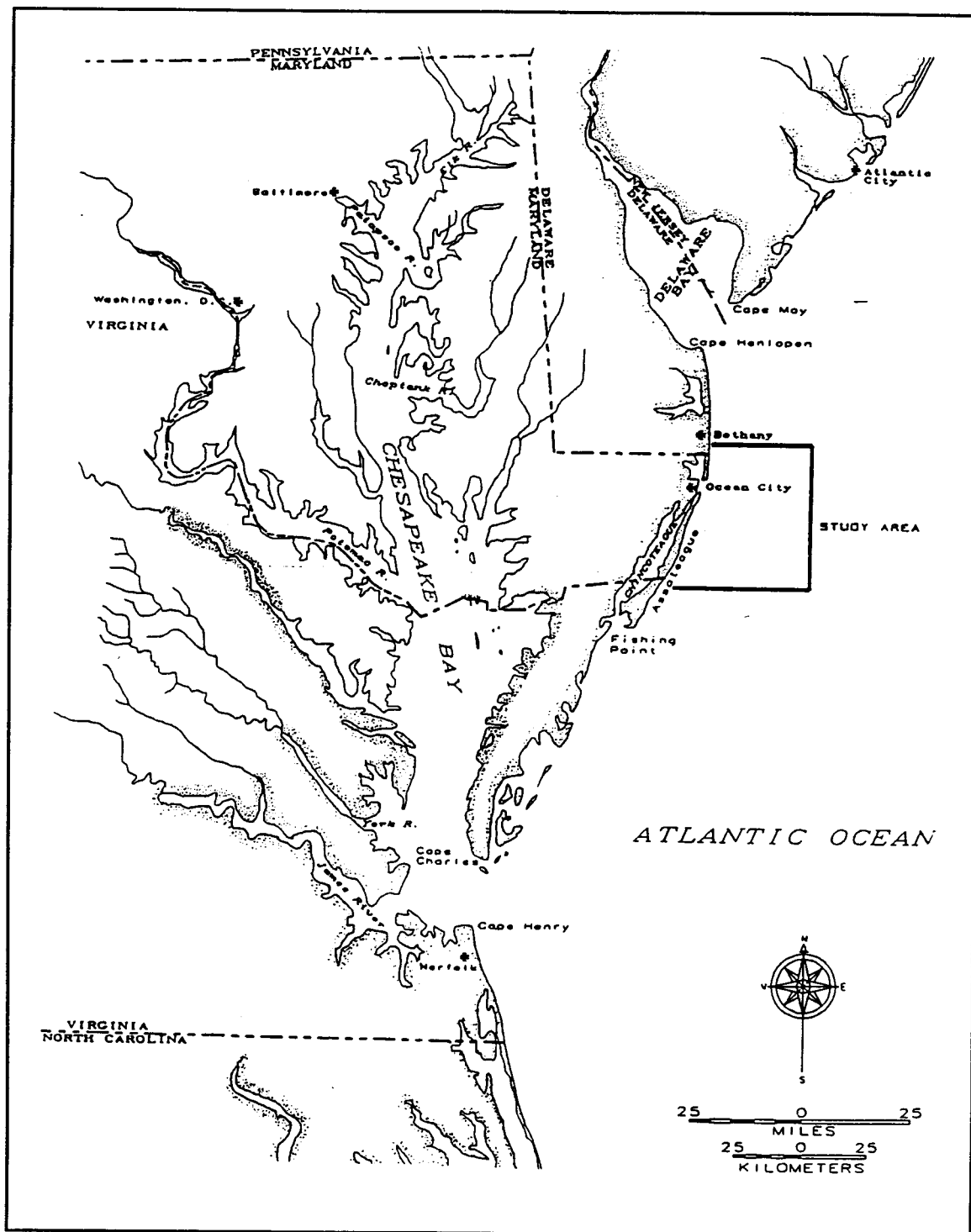


Figure 1. Location map of the study area.

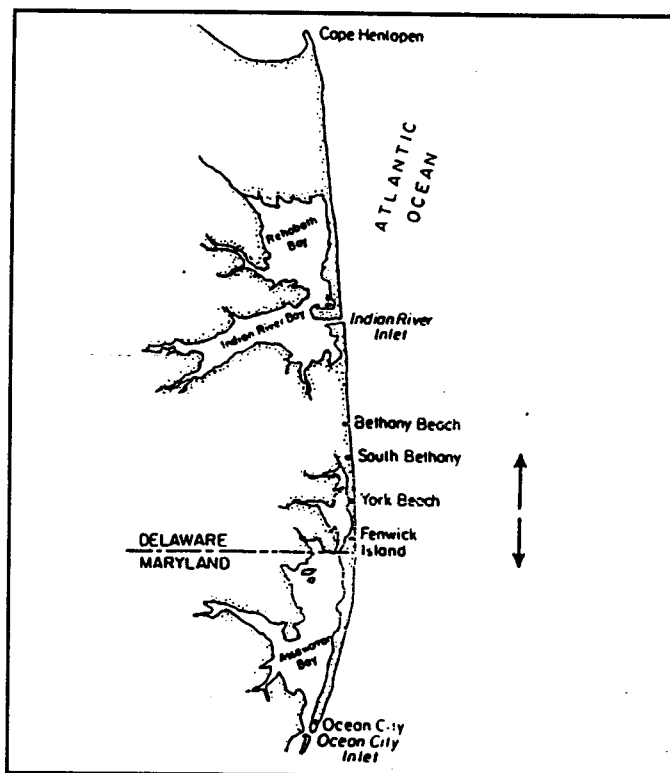


Figure 2. Location of divergent nodal point centered around York Beach (From Mann and Dalrymple, 1986).

growth of a series of recurved spits into Chincoteague Inlet since 1859 (Goettle 1978). Elevations along most of the barrier island are less than 3 m above MLW. The northern portion of Assateague Island, an 8.5 km stretch of shoreline south of Ocean City Inlet, is extremely narrow averaging between 120 to 215 m in width (Leatherman 1984). This portion is subject to frequent overwashing and has breached twice along the inside part of the south jetty. In November 1961, prior to the March 1962 Ash Wednesday storm, a breach was cut between the south jetty and the adjacent land (Figure 3). This breached area was exacerbated by the 1962 storm and required approximately .75 million m³ of sand to re-establish the shoreline. Neither of these two breach events were caused by hurricanes but by strong northeasters. In particular, the latter storm caused \$50 million dollars of property damages to the Delmarva coast (Truitt 1967).

Geological Framework

Ocean City and Assateague Island are within the physiographic region known as the Atlantic Coastal Plain. This is an eastward thickening wedge of partly consolidated/unconsolidated sediments, characterized by a surface of low relief (generally less than 61 m) and low gradient (generally less than 1.5 m per mile). These sediments range in age from Cretaceous, Tertiary, and Quaternary time periods (Halsey 1978). This area is deeply enbayed (i.e. Delaware and Chesapeake Bays), with the mainland coastal plain fringed by long continuous barriers. The Delmarva Peninsula owes much of its geomorphic state to glacial meltwater outwash derived from the northern ice fronts. The closest glacial ice fronts were the New Jersey-Pennsylvanian, located approximately 160 km to the north. Since the Wisconsin glaciation about 25,000 yrs B.P., this coastline has been undergoing transgression, as a result of rising sea level and dwindling sediment supply (Kraft 1971, Biggs 1970, Field 1979, Toscano et. al 1989, Halsey 1978). Evidence from vibracores and peat samples taken offshore from Ocean City and Assateague Island suggest the present shoreline has migrated landward from a position of at least 8 km seaward on the shelf (Biggs 1970, Field 1979, Halsey 1978). Preservation of these barriers is due in part to a natural process of barrier rollover. This landward movement of sediment buries backbarrier lagoonal environments, causing new marshes to develop along and/or on top of the overwashed sediment.

The Delmarva Peninsula lies within "the Mid-Atlantic Bight Coastal Compartment" as described by Fisher (1967 and 1982)(Figure 4). The Mid-Atlantic Bight was divided into five coastal compartments: Cape Cod, Massachusetts, Long Island, New Jersey, Delmarva (Delaware-Maryland-Virginia) and North Carolina Outer Banks. Each coast in this model exhibited similar morphologic characteristics: (1) a northern cusped spit, (2) an eroding headland, (3) a southern spit, (4) a barrier island chain (Figure 5). Fisher 1967, proposed that these barrier features originated along a shoreline of submergence rather than emergence (i.e. reworking of Pleistocene headlands and laterally prograding spits). Figure 6 shows the various geomorphic features within the Delmarva Coastal Compartment. The Mid-Atlantic Bight coastal compartment contains the vast majority (94 percent) of shoreface attached and detached sand ridges (linear shoals) off the continental United States (McBride and Moslow 1991). The local coastal processes affecting this

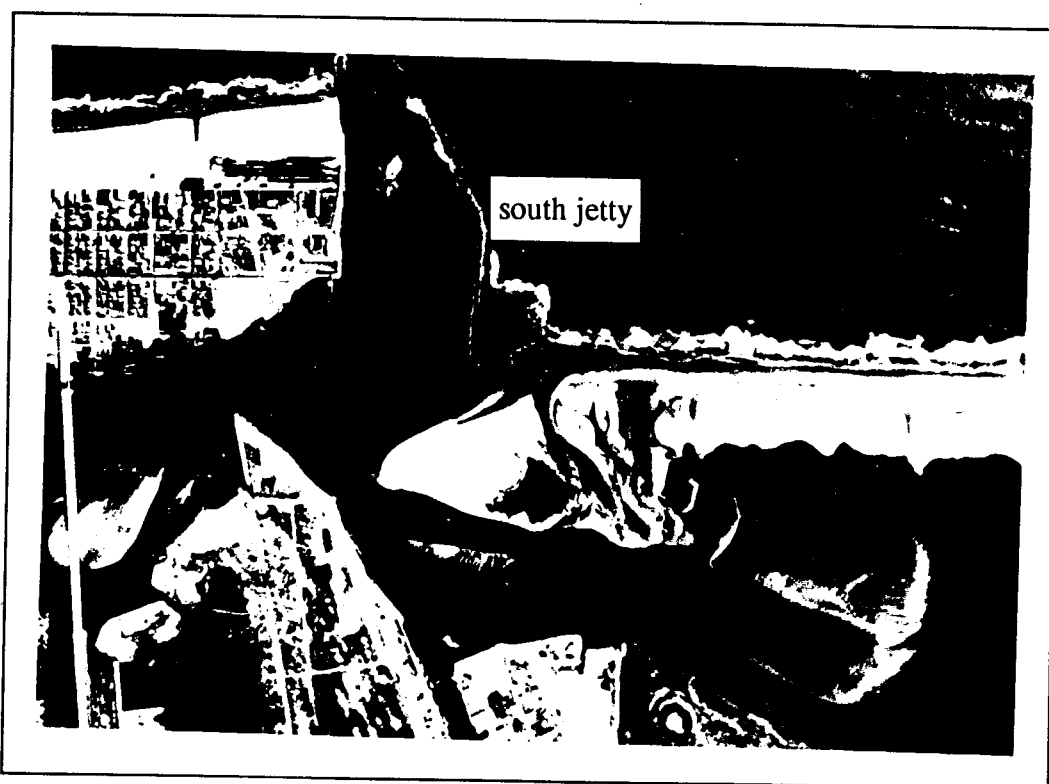


Figure 3. This photograph was taken, five months before March 1962 "Ash Wednesday storm", note the breach along the inshore portion of south jetty.

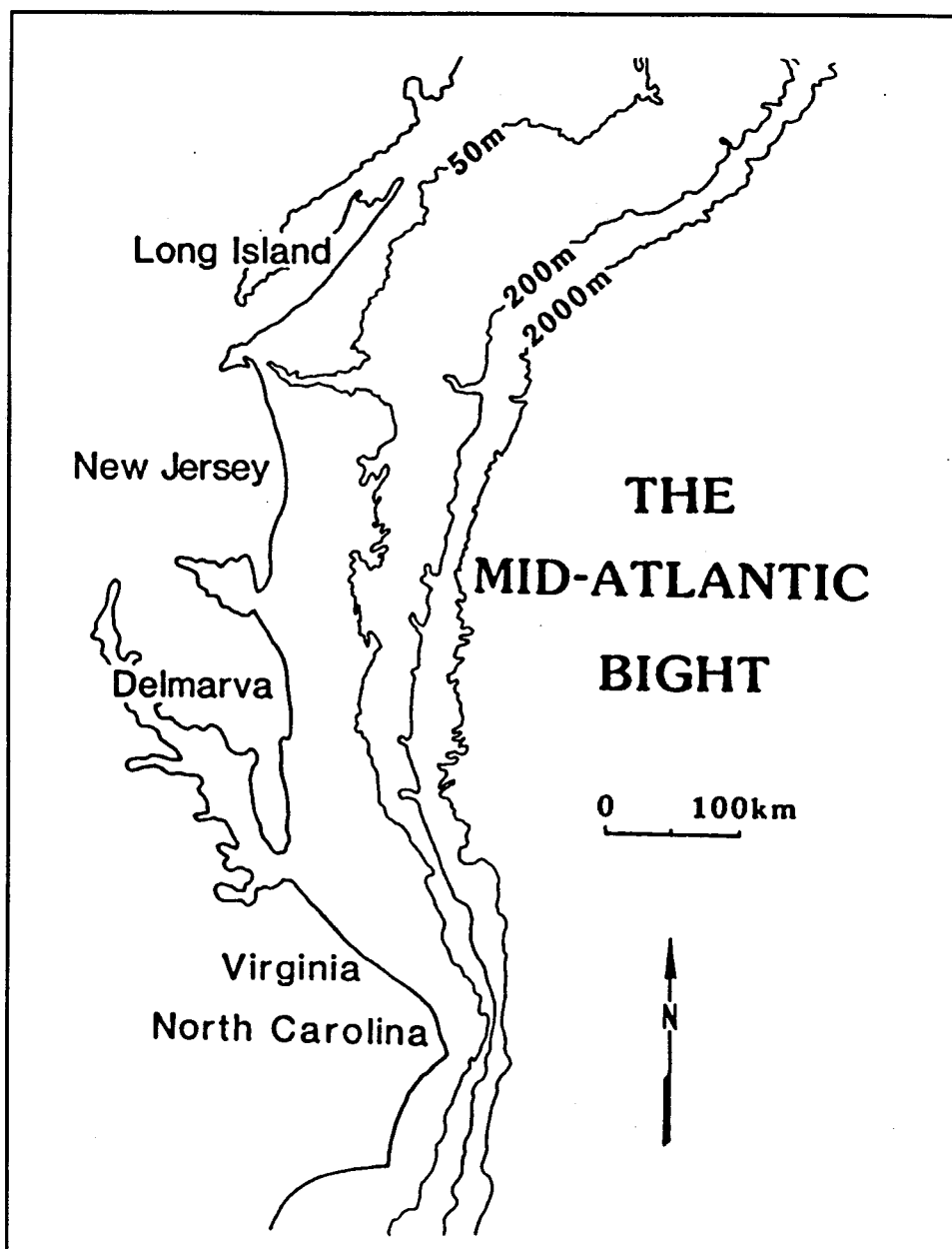


Figure 4. Location of coastal compartments within the Mid-Atlantic Bight (From McBride 1986).

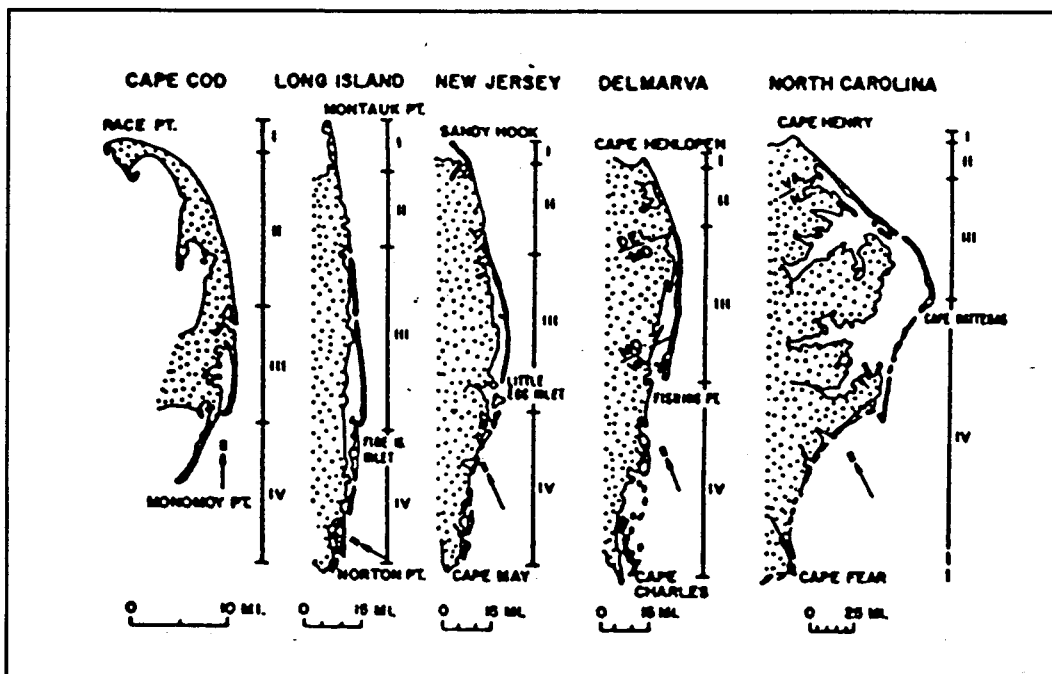
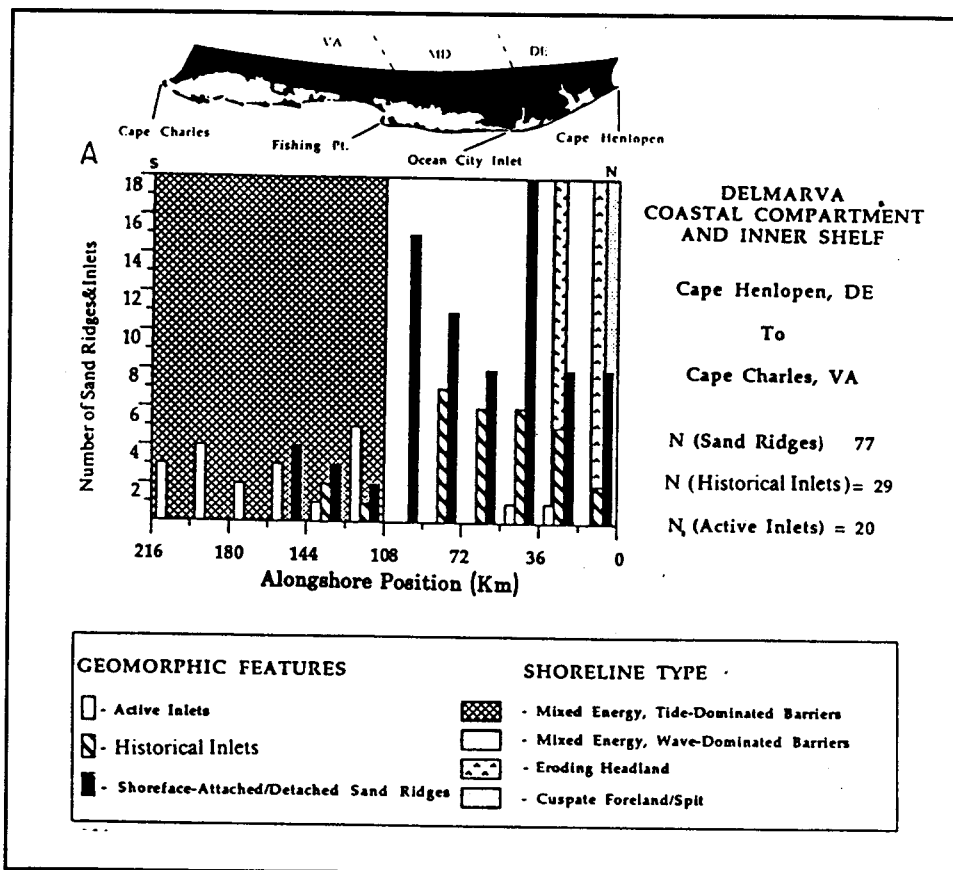


Figure 5. Coastal compartments of the Mid-Atlantic Bight (I) Northern or Cuspate Spit, (II) Eroding Headland, (III) Southern Spit, (IV) Barrier Island Chain (After Fisher, From Swift 1969).



region provide ideal conditions for formation of these linear sandbodies: (1) barrier transgression, (2) mixed energy, wave dominated shorelines, (3) and lateral inlet migration in a southerly or southwesterly direction. McBride (1986) mapped approximately 54, 71, 77, and 43 sand ridges offshore of the Long Island, New Jersey, Delmarva, and Virginia/North Carolina coastal compartment, respectively. There are a number of common attributes of these features. They are generally oriented at oblique angles to the shoreline (average orientation 29°), composed of unconsolidated fine to medium grained sand, generally over 1000 m long, have relief to 10 m, have average side slopes of 1° , are 1-3 km wide with wave lengths of 1-8 km. The origin of these features have been postulated to be from former ebb-tidal deltas. McBride and Moslow (1991), presented a six stage model of development for these offshore sand bodies (Figure 7). The sequence begins with a shoreline irregularity (sediment accumulation - relict ebb tidal shoal ?). The shoreline begins to transgress, the shoreface connected shoal becomes elongated to the direction of littoral drift and landward barrier migration. Eventually, detachment occurs and an offshore shoal is left behind. Fenwick and Assateague Islands have the largest number (63) and highest density of attached and detached sand ridges (.73 ridges/km) (McBride and Moslow 1991). These offshore shoals have also been evaluated for their suitability and compatibility for beach nourishment, and consequently were used for beach replenishment along Ocean City's and Delaware's shoreline (Anders and Hansen 1990).

The Delmarva Peninsula's most dominant features are its barrier islands found along its eastern margin, bordering the Atlantic Ocean. This stretch of shoreline (approximately 190 km in length) from Cape Henlopen, De to Cape Charles, Va, exhibits obvious morphological differences in the northern and southern regions of this coast. The northern barriers tend to be long and continuous, while the southern barriers are short, segmented islands, with numerous permanent natural inlets. Truitt (1968), and McBride (1986), have documented and referenced various locations of transient inlets along Fenwick and Assateague Islands. These inlets tended to be ephemeral in nature, originating during storm conditions, eventually closing due to a combination of natural coastal processes and sediment supply. The stretch of shoreline from Cape Henlopen De, to Chincoteague Inlet, Va, has been the location of numerous tidal inlets (26), of which only 2 remain presently active (McBride 1986). The two active inlets are Indian River Inlet, De, which was artificially cut in 1929, and Ocean City Inlet, Md. Both of these active inlets, were artificially stabilized in order to remain permanently open. Southern Delmarva, from Chincoteague Inlet, Va to Cape Charles, Va, has been the location of 23 tidal inlets, of which 18 are presently active (McBride and Moslow 1991). Tidal inlet frequency and barrier island length have also been linked to differences in tidal range between the northern and southern portions of the Delmarva Peninsula (FitzGerald 1988) (Figure 8). As tidal range increases, tidal prisms also increase, necessitating a greater number of tidal inlets to accommodate the larger volume of water flowing into and out of the backbarrier lagoons. Table 1 contrasts differences in the northern and southern regions of the Delmarva Peninsula.

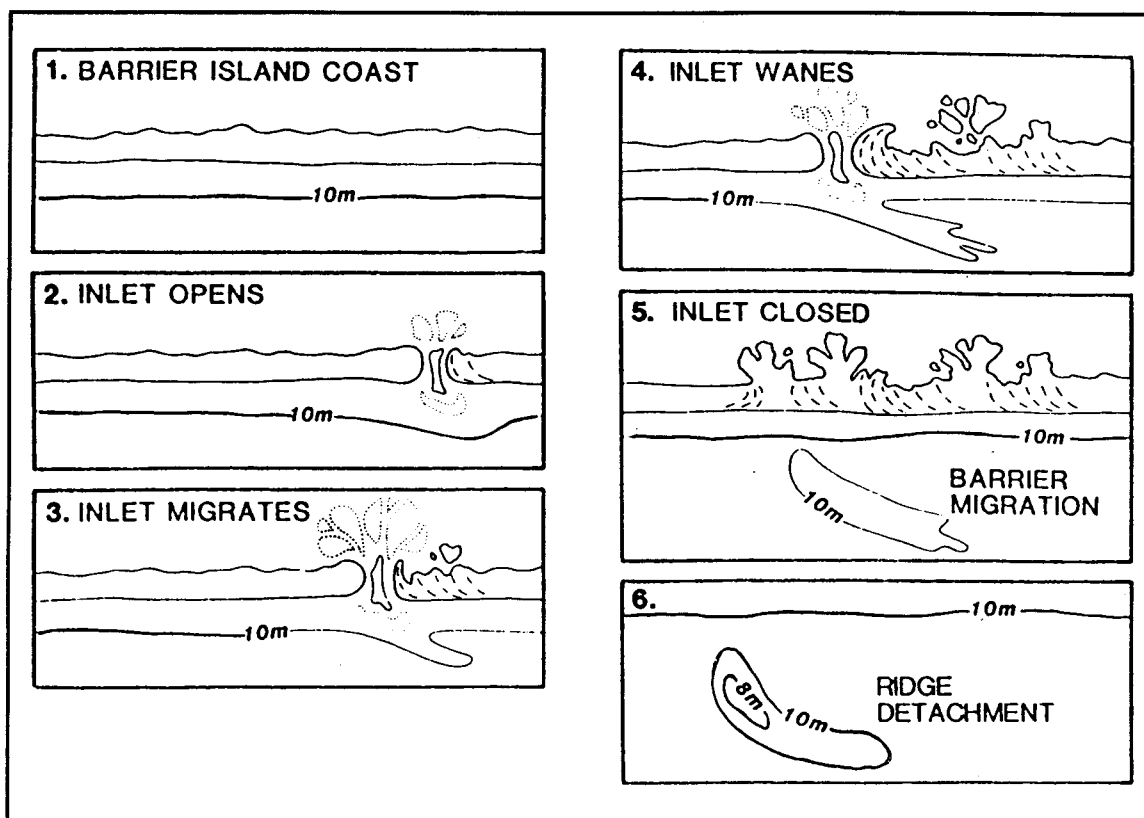


Figure 7. Six stage ebb-tidal model illustrating the importance of tidal inlet and barrier island morphodynamics from the origin, evolution and distribution of shoreface attached sand ridges (from McBride and Moslow 1991).

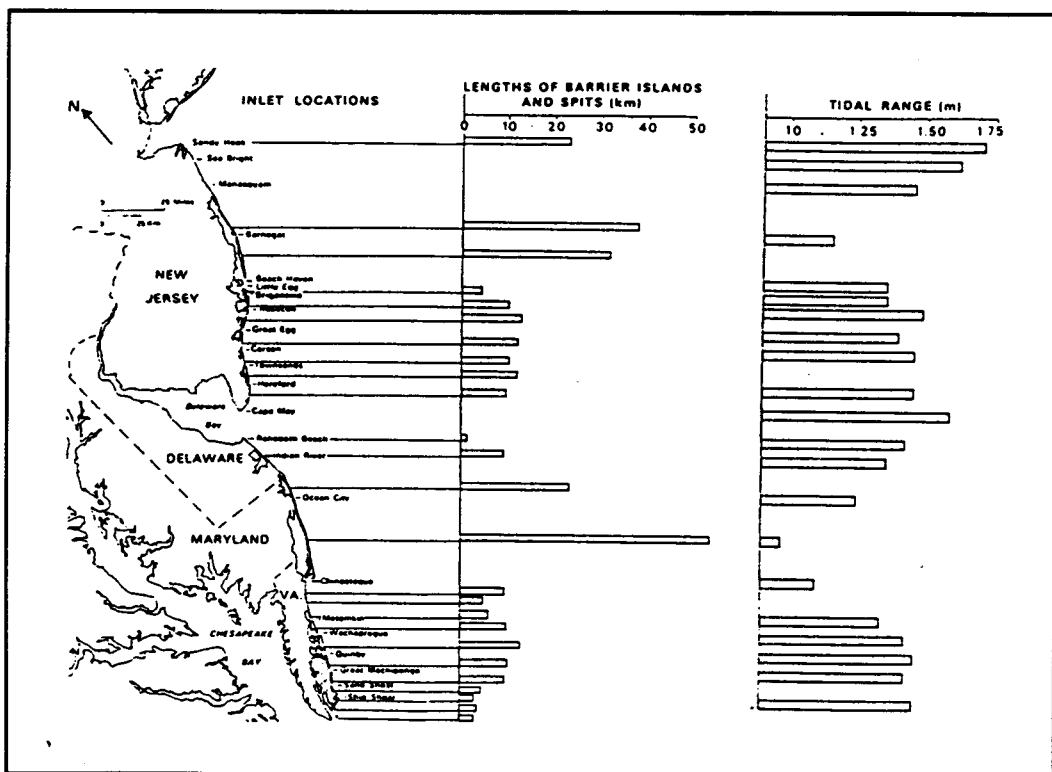


Table 1 Variation in northern and southern regions of the Delmarva Peninsula (Modified from Biggs 1970, McBride 1986, McBride and Moslow 1991, and Swift 1975)		
Description	Northern	Southern
Offshore Slope	Increased	Decreased
Primary Sediment Source	Eroding Headland	Offshore
Avg/Maximum Barrier Width	3.1 km	3.39 km
Avg/Maximum Barrier Length	44.5 km	8.18 km
Grain size	Coarser (.88 mm)	Finer (.62 mm)
Shoreline Type	Mixed energy Wave dominated	Mixed energy Tide dominated
Sand Ridge Development	Greater (69%)	Lesser (22%)

Coastal Processes

The Ocean City/Assateague Island coastline is classified as microtidal (0-2 m) (Davies 1964), and has a semidiurnal tide with a mean range of 1.0 m. The extreme range (spring tide) is approximately 4 m, fluctuating from 1 m below to 3 m above MLW (Field 1979). Average air temperature is 56°F (13.3°C) and mean annual precipitation is 119.4 cm (Mather, 1969). In the vicinity of Ocean City and Assateague Island, the estimated annual net littoral drift ranges between 114,750 m³ and 214,200 m³ to the south; although the direction of littoral drift changes periodically (Dean and Perlin 1977, Douglas 1985, Mann and Dalrymple 1986) (Figure 9 and 10). Figures 11 and 12 show wind and wave information (Hubertz et al 1993). For the 20 year period 1956 to 1975, average wave height (H_{mo}) in 16 m of water was 1.04 m, with maximum wave height of 7.5 m, peak period (T_p) was 15 sec and peak direction (D_p) was recorded at 68 deg. The predominate wave period ranged between 5 and 7 seconds. Approximately 22 percent of waves along this segment of coast approach from the southeastern quadrant and 95 percent of the time these waves are less than 2.4 m high. Predominate southerly littoral drift is a result of waves from the northeast and eastern quadrant 5 and 29 percent of the time, respectively. Waves in excess of 2.4 m occur from the northeast and east 10 percent of the time. Winds from the northeast average above 30 knots, 10 percent of the time, and winds from the east average 25 knots, 5 percent of the time. Wind speeds from the southeast are less than 20 knots, 90 percent of the time. Wave Information Study (WIS) data of hindcast wave heights and maximum wave height averaged for the period 1956 to 1975 are shown in Figure 13. Average maximum wave heights are greater throughout the fall and winter months (September-March), decreasing through spring and summer

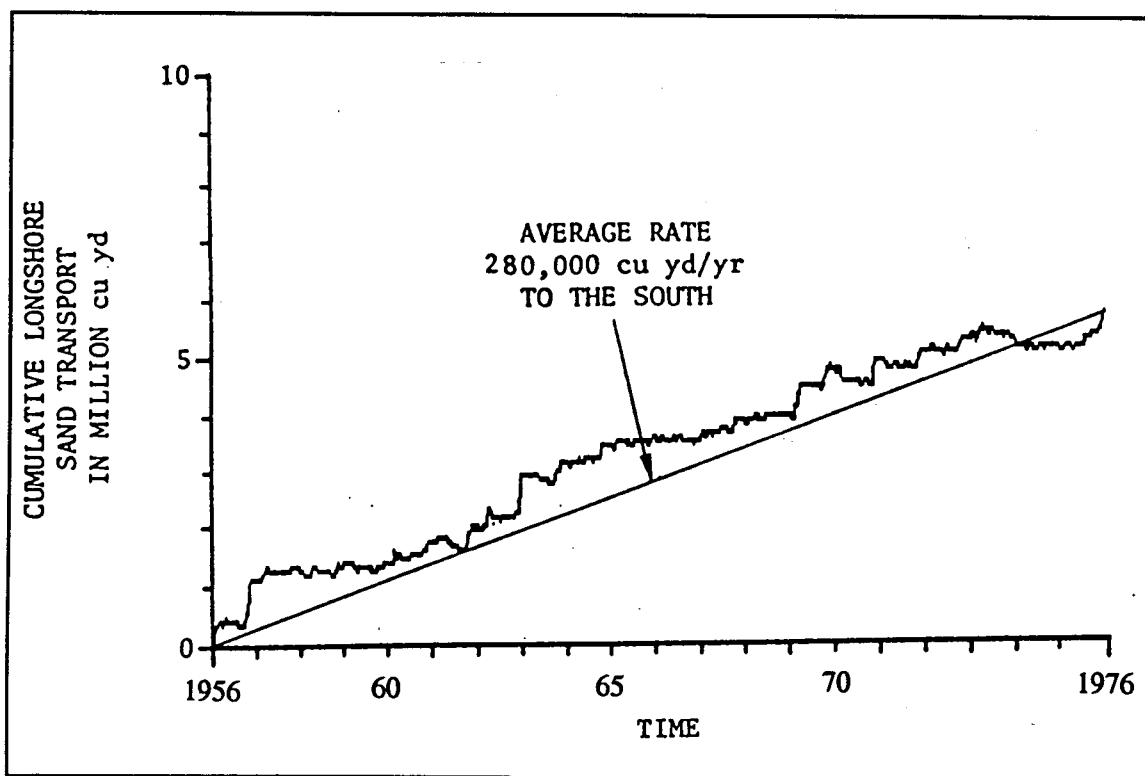


Figure 9. Net average southerly drift for Ocean City, Maryland, from hindcast data 1956-1975 (From Douglas 1985).

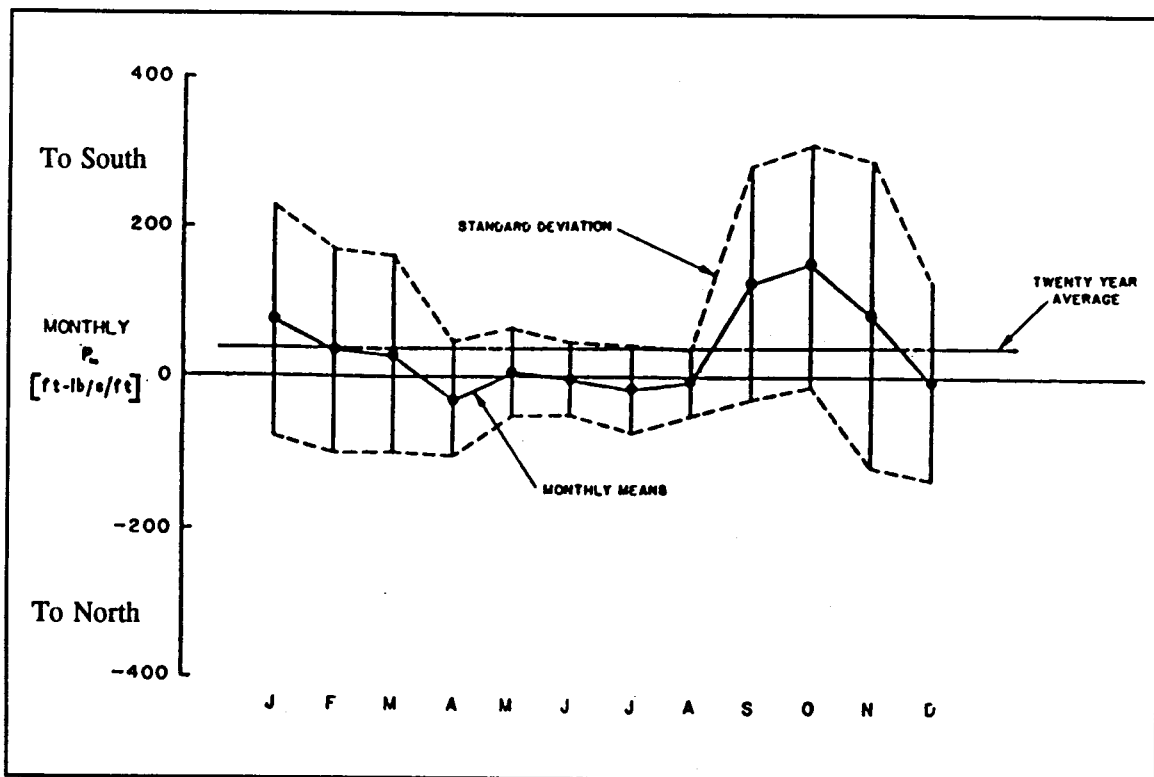


Figure 10. Seasonal fluctuations of littoral drift for Ocean City, Maryland from hindcast data, 1956-1975 (From Douglas 1985). Southerly directions are above zero, northerly directions are below zero.

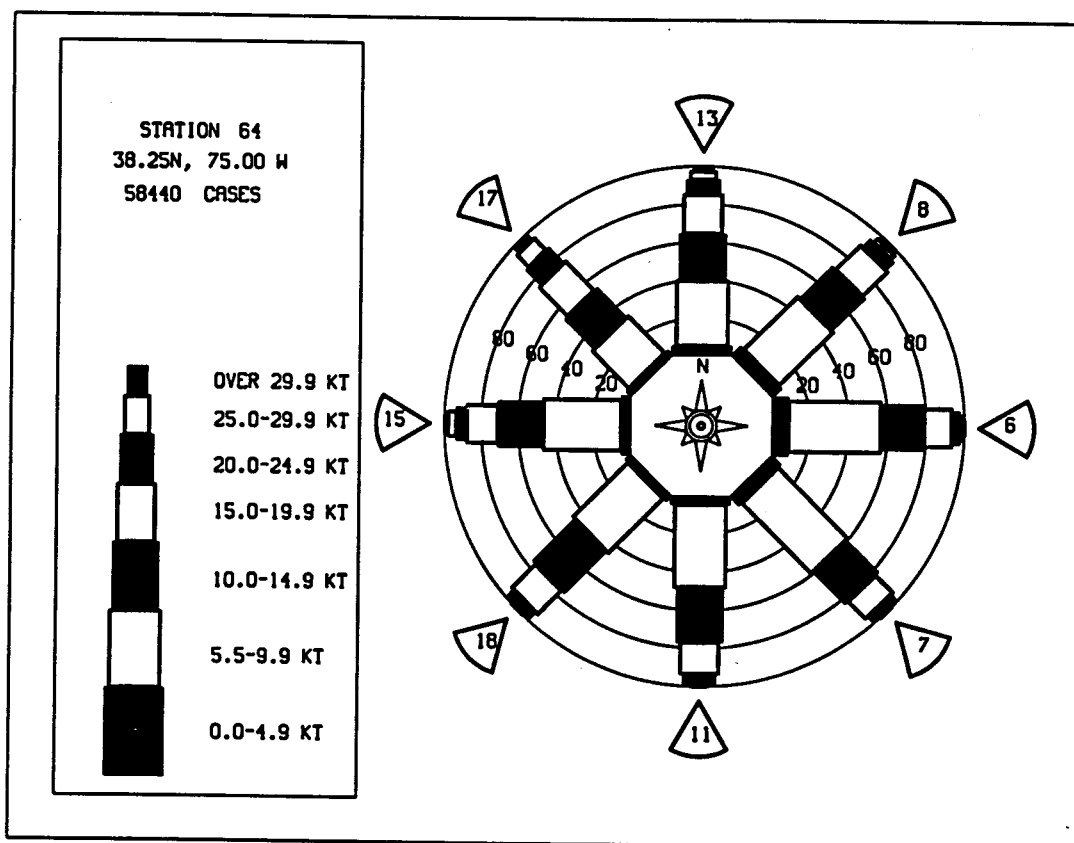


Figure 11. Wave Information Study (WIS) hindcast wind rose, from 1956-1975 for station 64 offshore Ocean City, Maryland (Hubertz et al 1993).

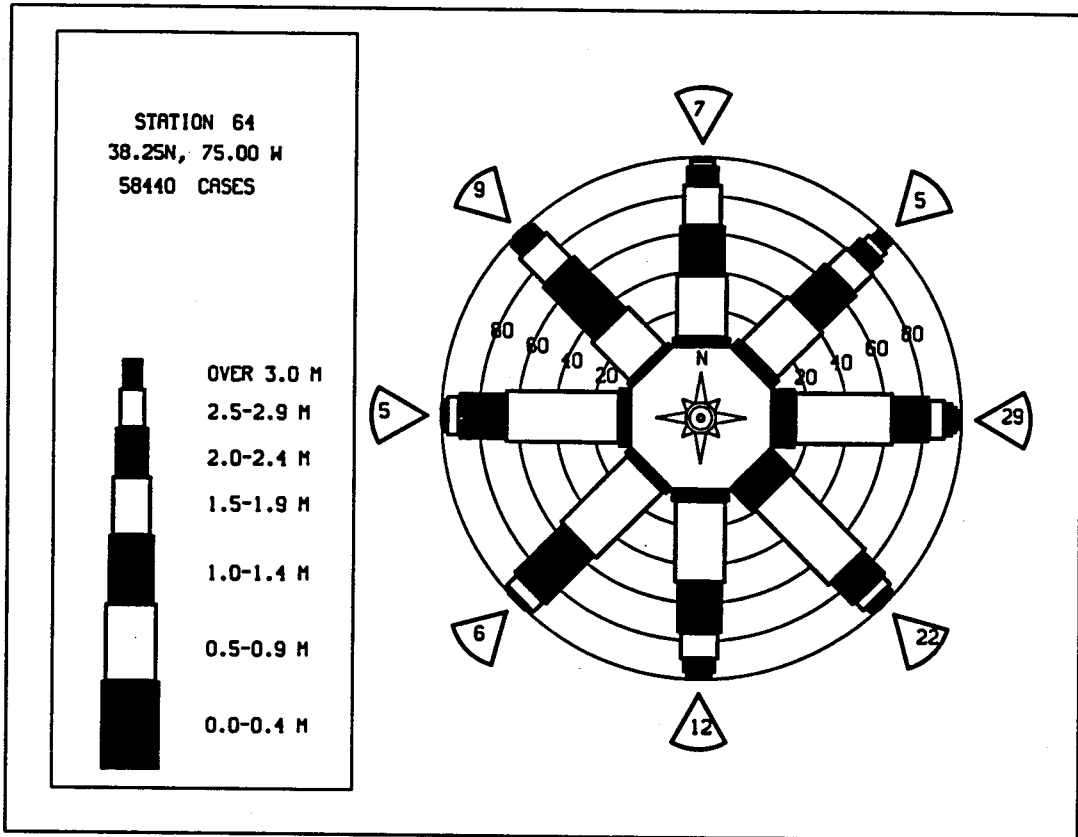


Figure 12. Wave Information Study (WIS) hindcast wave rose, from 1956-1975 for station 64 offshore Ocean City, Maryland (Hubertz et al 1993).

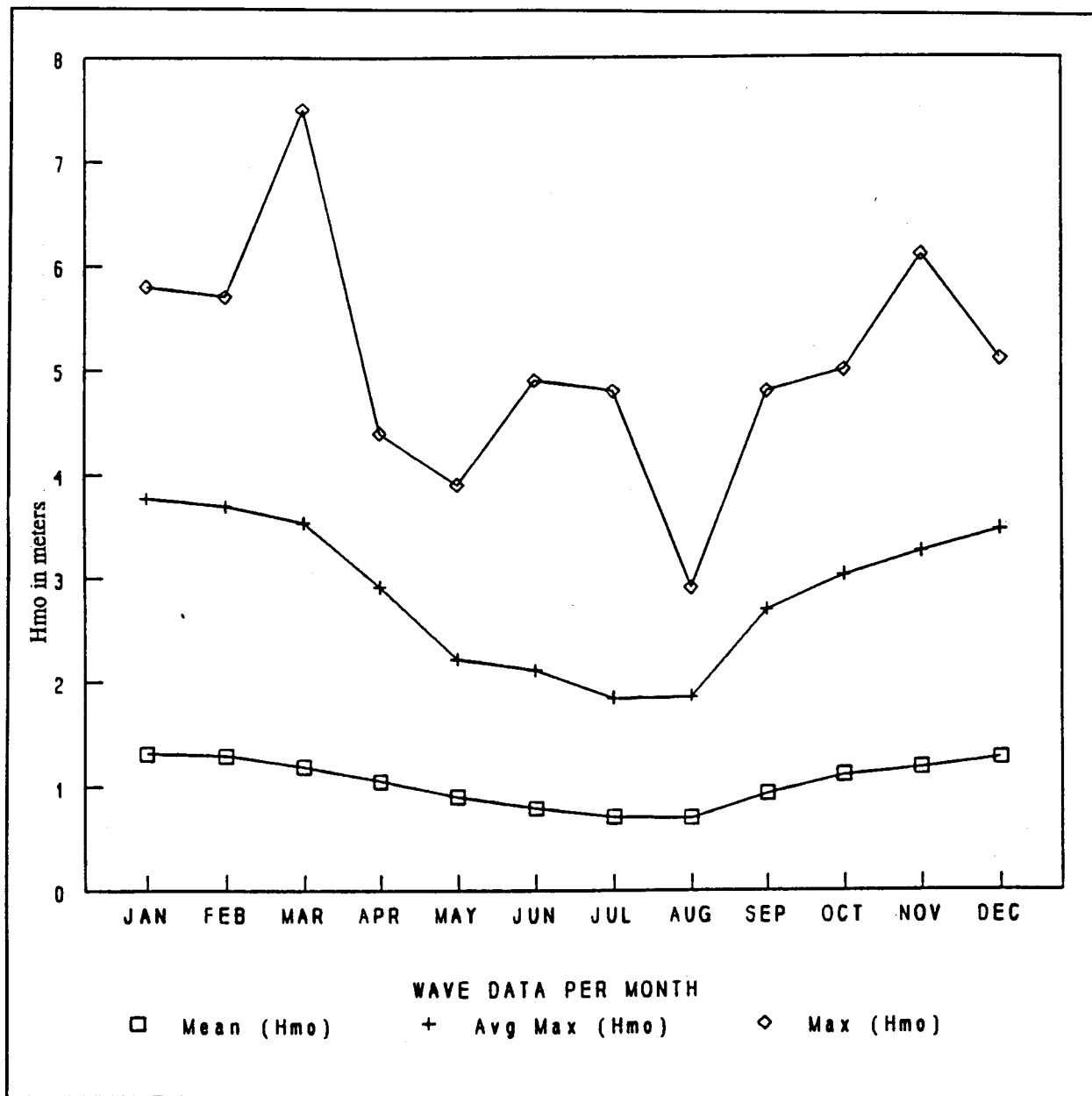


Figure 13. Hindcast (mean, avg. max, and max Hmo) from hindcast data 1956-1975 for station 64 offshore Ocean City, Maryland (Hubertz et al 1993).

(April-August). For the 20 year period 1956 to 1975, the largest maximum wave height occurred in March, and the smallest maximum wave height occurred in August.

Storm Events

Tropical storms (hurricanes) and extratropical storms (northeasters) are not uncommon occurrences along the Delmarva Peninsula. The Atlantic hurricane season extends from June 1 through November 30. The origination or breeding grounds for East Coast hurricanes is predominately off the western coast of North Africa, within 10 degrees of the equator. Historically, Maryland has not been on the mainline of hurricane travel; most tropical hurricanes are deflected either east or west (offshore or onshore) of the study area. Since 1900, the most destructive storms recorded that effected the study area were the 1933 hurricane that created Ocean City Inlet, hurricane Hazel in 1954, hurricanes Connie and Diane in 1955, hurricane Donna in 1960, the "Ash Wednesday" extratropical storm in 1962 (Truitt 1967), and the January 1992 extratropical storm (Grosskopf and Behnke 1993). The tracks and origination points of the hurricanes are shown in Figure 14. Table 2 shows a complete listing of tropical storms and their associated classification, that have passed within a 185 km radius of Ocean City from 1899 through 1992. Extratropical storms (northeasters) usually have milder wind velocities (less than 75 mph from the NE quadrant) but tend to be of longer duration than hurricanes. Extratropical storms are more common from December to April, with very few between June and August (Figure 15) (Dolan and Davis 1992). There are two common types of strong northeasters that effect the study area: Bahama and Florida Lows (Figure 16 and 17). Bahama Lows are common in October, formation occurs north of Cuba, intensifying over the open ocean in a slow northern track. Florida Lows are common in March, developing near the Florida peninsula, and track in a north-northeast direction. In general, a northeasters destructive power is derived through a combination of size, intensity, and speed of the storm, tidal phase (spring or neap), storm path with respect to the beach, and time interval between storms (Hayes and Boothroyd 1969). Dolan et al (1988) tabulated a total of 1,347 Atlantic northeasters from 1942-1984. Storms with recorded deep water wave heights of 1.5 m were examined and classified into different categories based on storm frequency and magnitude. This 1.5 m cutoff was based on heights needed to produce minimal coastal erosion along North Carolina's barrier islands. Dolan and Davis (1992) modified the Sapher-Simpson Scale for tropical cyclones, and developed five new classes to describe a storms "destructive power". This power function was related to maximum wave heights and storm duration. These classes and their frequency of occurrence are shown in Figure 15.

From October 1991 through January 1992, three very significant "northeasters" had a major impact on Ocean City and Assateague Island. Table 3 depicts associated storm process information from these three storms. The first northeaster, dubbed the "Halloween Storm", occurred between 29 October - 1 November. This four day storm was recorded as producing maximum offshore and nearshore wave heights of 5.0 m and 3.1 m respectively (Jensen and Garcia 1993). Nearshore NOS tide gauges at Ocean City recorded peak water levels of 1.57 m (Jensen and Garcia 1993).

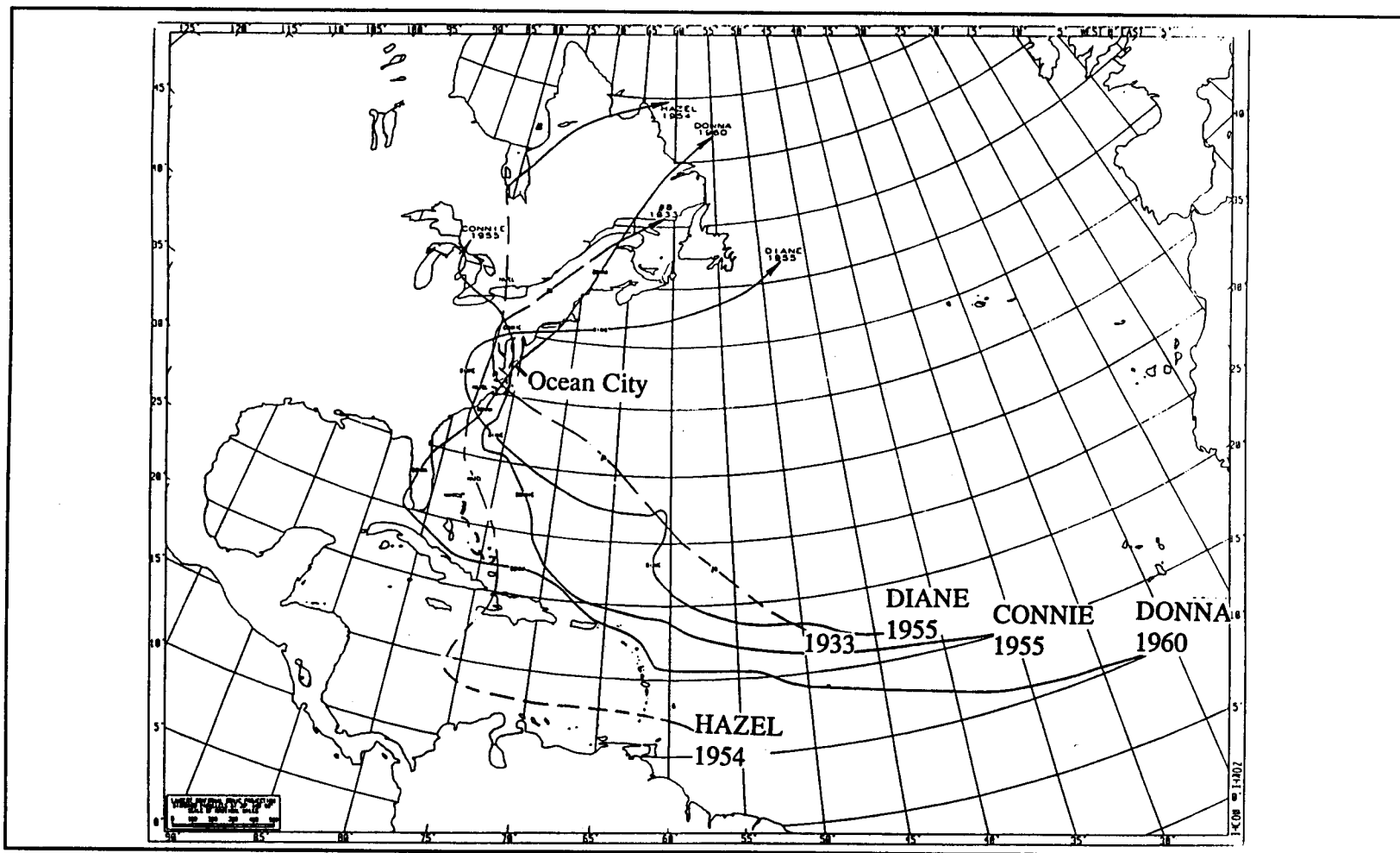


Figure 14. Hurricane and origination points for storms in 1933, 1954, 1955, and 1960 (Modified from Neuman et al 1992).

Table 2

Total number of tropical storms that passed within 92.5 km and 185 km of Ocean City Inlet from 1899-present. Modified from Neuman et al 1992.

YEARS	TOTAL STORMS		¹ STAGE OF DEVELOPMENT									
	< 92.5 km (a)	< 185 km (b)	^a TDdev		^b TS		^c HU		^d TDdis		^e EX	
			a	b	a	b	a	b	a	b	a	b
1899-1910	7	4					1	2			6	2
1911-1920	1	2			1	1				1		
1921-1930	4	3			1	1		1			3	1
1931-1940	6	4			1	1	3			1	2	2
1941-1950	5	5			2	3	1				2	2
1951-1960	7	5			3	1	3	2			1	2
1961-1970	5	5	1	1	2	2		1	1		1	1
1971-1980	2	3			1	1		2	1			
1981-1990	5	2			2	2	2				1	
1991-1992	0	1						1				
TOTALS	42	34	1	1	13	12	10	9	2	2	16	10

¹ Classification criteria for Tropical Cyclones from 1899-present.

Tropical Depression (development) used from 1951-1986, The formative stages of tropical cyclone, maximum sustained surface wind is < 34 kt.

^b Tropical Storm used from 1899-1986, A warm core tropical cyclone, maximum sustained surface wind ranges from 34 to < 64 kt.

^c Hurricane used from 1899-1986, A warm core tropical cyclone, maximum sustained surface wind is > 64 kt.

^d Tropical Depression (dissipation) used from 1899-1986, The decaying stages of a tropical cyclone, maximum sustained surface wind is < 34 kt.

^e Extratropical Cyclone used from 1899-1986, Tropical cyclones modified by interaction with nontropical environment. There is no wind speed criteria and maximum winds may exceed hurricane force.

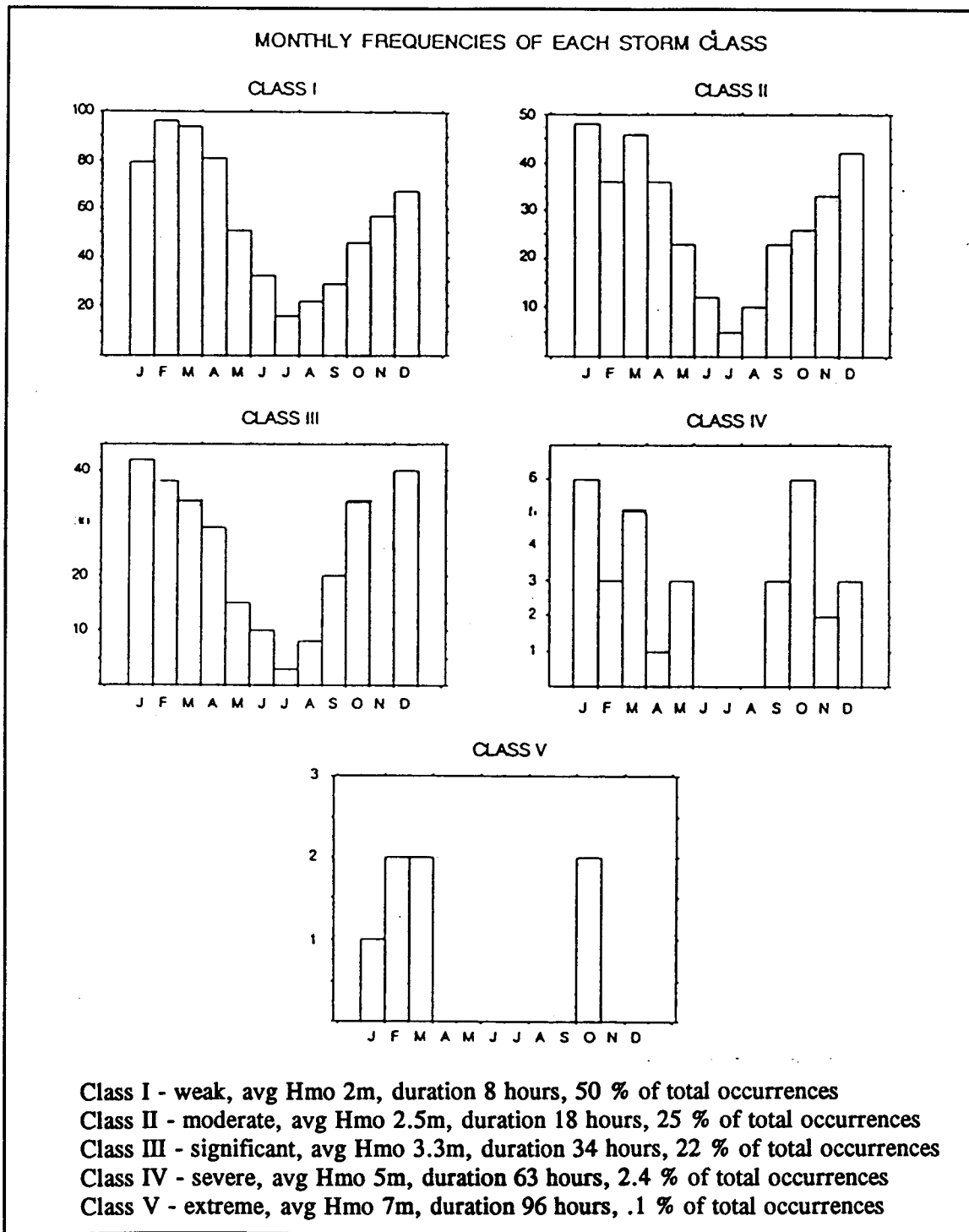


Figure 15. Monthly frequency of extratropical storms (From Dolan and Davis 1992).

BAHAMAS LOW

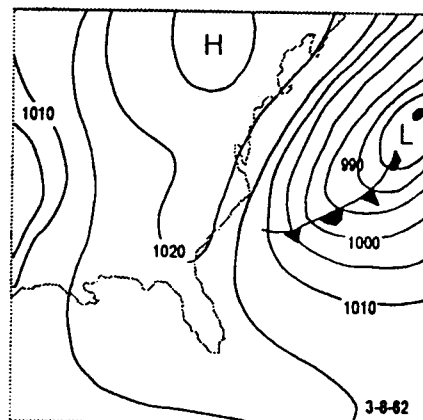
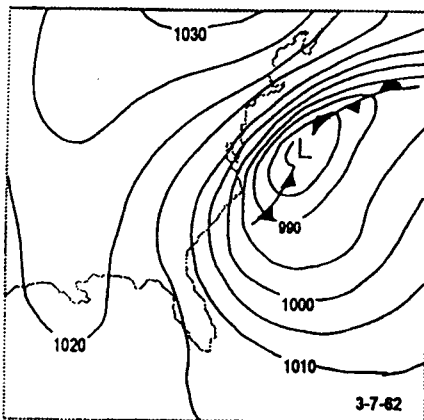
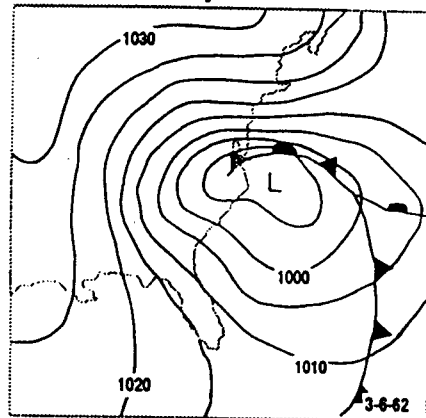
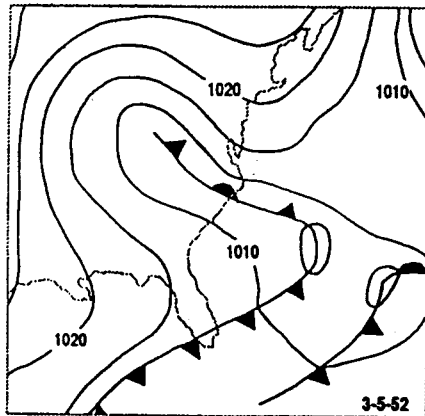


Figure 16. An example of a strong northeaster that forms near the Bahamas (From Dolan and Davis 1992).

FLORIDA LOW

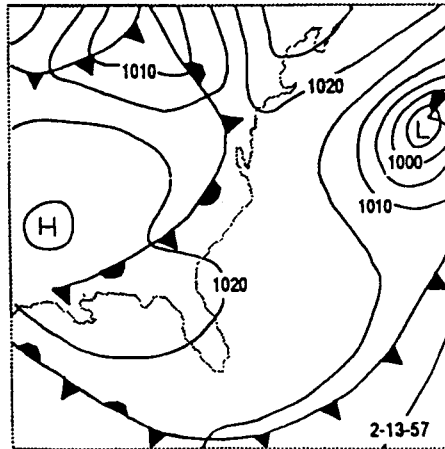
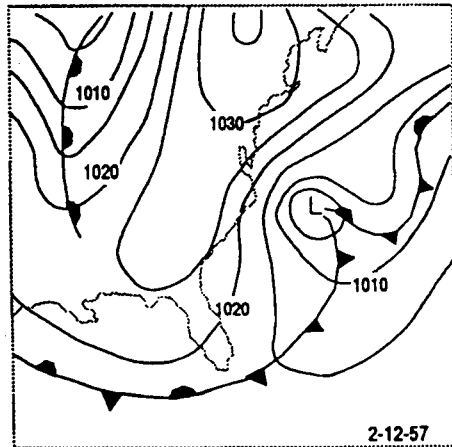
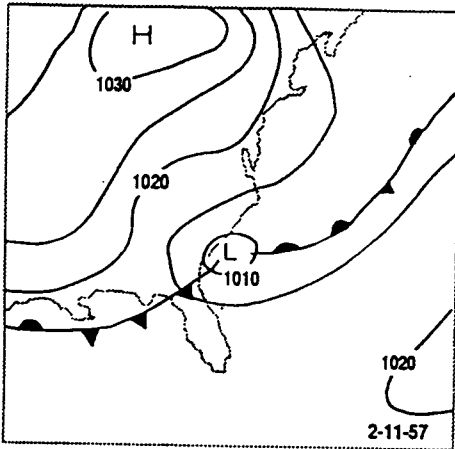


Figure 17. An example of a strong northeaster that forms over the Florida peninsula (From Dolan and Davis 1992).

Table 3 Comparison of three recent "northeaster" storms from October 1991 to January 1992. Modified from Kraus and Wise 1993.			
Factor	Halloween Storm 1991	November 11, 1991	January 4, 1992
Significant Wave Height	(8.0 m)	(3.0 m)	(8.5 m)
Wave Period, sec	25.0	10.0	14.3
Surge near Shoreline	(1.6 m)	(1.2 m)	(2.0 m)
Duration, hr	96	48	24
Wind Direction	Alongshore	Onshore and Alongshore	Onshore
Wind Speed	24 kt (12.3 m/sec)	20 kt (10.3 m/sec)	35 kt (18.0 m/sec)

Dolan and Davis 1992, characterize this storm as the most powerful northeaster to affect the Atlantic Coast, (i.e. measured wave height), surpassing the 1962 "Ash Wednesday" storm. Fortunately, the storm assumed a more easterly direction, allowing its more destructive forces to be felt north and east of Ocean City. The second northeaster occurred on November 11, 1991. This storm location was more localized than the Halloween storm, bringing high winds and waves, compounding damages that had occurred from the Halloween storm approximately two weeks earlier. Then on January 4, 1992, the study area experienced a more localized storm positioned just offshore of the Delmarva peninsula. Storms water levels at Ocean City were only +.37 m lower than reported levels associated with the "Ash Wednesday" storm (Grosskopf and Behnke 1993). What made this storm so devastating was its position in relation to Ocean City, and its direct onshore wind patterns elevating nearshore waves, and coastal water levels (Jensen and Garcia 1993).

The influence of storms (both tropical and extratropical) on longshore sediment transport at Ocean City is shown in Figure 18. Approximately 50 percent of the total longshore sediment transport occurs 5 percent of the time, 67 percent of the longshore transport occurs approximately 10 percent of the time, and 90 percent during 35 percent of the days (Douglas 1985). This data emphasizes how a few storm events account for almost all longshore sediment transport at Ocean City.

Cultural History

Assateague is an Indian name meaning "Across" or "Place Across", implying its location in relation to the mainland. The early inhabitants were the Pocomoke, Chincoteague, and Assateague Indians of the Nanticoke tribe. It is known that Indians frequented the island and camped on it as early as or earlier than the 1600's. Many Indian artifacts have been found in the western portion of Ocean City, including a large burial ground located near the entrance of the first state highway bridge that linked the barrier island to the mainland. Encroachment by English colonists in the late 1600's, resulted in migration of these Indian tribes northward through Pennsylvania, New York and Canada (Hurley and Hurley 1979). Through the early nineteenth century, English settlements existed on Assateague Island, in particular North Beach village, located approximately ten miles south of Ocean City. Most settlements utilized the area for agricultural purposes (gardening) and for grazing of livestock. The southern portions of Assateague were preferred over the northern portions because of higher ground and wider expanses of land. This area was very rich in wildlife, waterfowl, and seafood resources. In July 1876, Ocean City, Maryland was officially named, and by 1893, coal burning locomotives were arriving daily from the mainland (Hurley and Hurley 1979). The railroad was a significant factor in public acceptance of Ocean City. Employment with the railroad, and other related skilled crafts, summer tourism and the pound fishing industry increased the population and stability of the fledgling town. The early 1900's brought the era of the great hotels which had a significant influence on the character of Ocean City. A severe

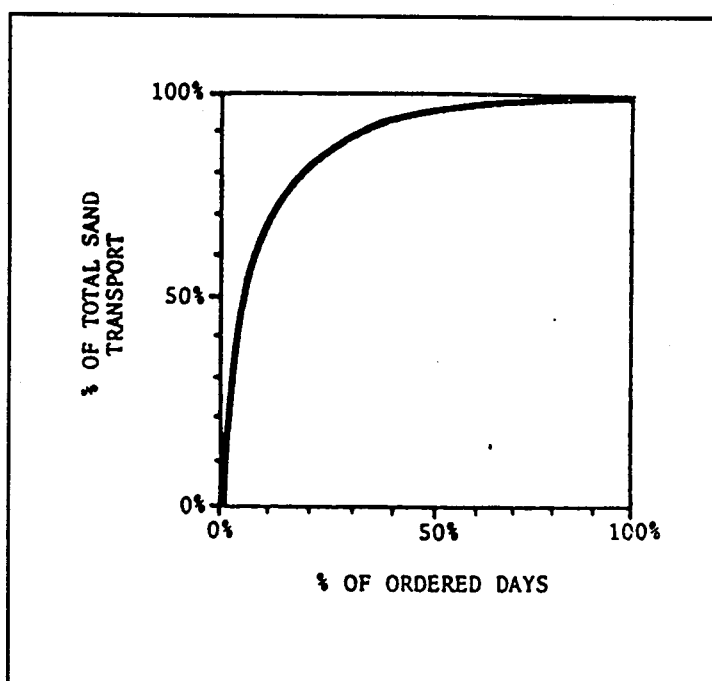


Figure 18. The influence of storms (both tropical and extratropical) on longshore sediment transport, from hindcast data (From Douglas 1985).

northeaster in February 1920 opened an inlet approximately 4.8 km south of Ocean City (Figure 19). This inlet migrated laterally southward for about .8 km, and eventually was closed by a northeaster on May 9, 1928. On August 23, 1933 another inlet was cut through what was then South Second Street and Third Street (Figure 20). The new inlet was created by breaching of Fenwick/Assateague spit, scouring the bottom to a depth that was almost immediately navigable to commercial vessels (approximately 3 m). This storm changed the face and life-style of Ocean City. Commercial fishing camps that dominated the area were leveled and the railroad bridge linking Ocean City to the mainland was destroyed. Four days after the inlet was breached, Captain D. Frank Parsons was the first to take paying guests through the inlet to go fishing offshore, little realizing that a whole new era of fishing was beginning for the town. During the season of 1934, two brothers from Selbyville, Delaware, Jack and Paul Townsend, pioneered the billfishing sport by chartering a vessel to search for marlin. It is said that commercial fisherman had often reported seeing large fish "tailing" offshore. By 1939, local sportfishing captains had learned to hook these fish quite well, and the city was being referred to as the White Marlin Capital of the World. The first blue marlin was caught in 1941, a second one the next day. The third was not boated until 1953. Today sportfishing is a primary industry in the fabric of Ocean City's economy. After World War II along Fenwick Island, commercial development increased, and in 1944, construction of a new access bridge was completed. In the late 1970's, commercial development peaked with construction of numerous high rise condominiums. Currently, Ocean City is recognized as Maryland's principal seaside resort. The permanent population is still less than 6,000, but boasts a summer population exceeding 250,000 (Leatherman 1987b).

In 1935, the National Park Service inventoried all unspoiled coastal areas along the Atlantic Coast, and recommended Assateague Island for federal protection. However, Congress took no action for special protection for Assateague Island. In 1943, Chincoteague National Wildlife Refuge (located on the Virginia portion of Assateague) was established through funding from sale of duck stamps to hunters (National Park Service 1985). By 1955, the National Park Service concluded that Assateague was already lost to private development and ownership and did not recommend the Maryland portion for national seashore designation. During the March 1962 northeaster, approximately 32 of the 50 dwellings located on Assateague were totally lost, and seven others were severely damaged. Private interests requested federal assistance to implement shore protection measures for prevention of any future damage caused by storms. Cost to benefit ratios were extremely high, resulting in rejection from the Federal government for any monetary assistance.

During the early 60's, additional federally protected coastal recreation areas were needed due to rapidly expanding urban populations along the eastern seaboard. In 1963, a meeting was held by the Interior Secretary's non-governmental Advisory Board on National Parks, Historic Sites, Buildings and Monuments. The committee endorsed the proposal to create Assateague Island National Seashore, but Congress adjourned without finalizing any formal legislative action. In 1965, the Assateague Island National Seashore was officially authorized. During this period, a portion of island south of the Sinepuxent bridge was already inhabited and owned by approximately 3,000 individuals (5,850 lots). A provision was inserted in the 1965 agreement for construction

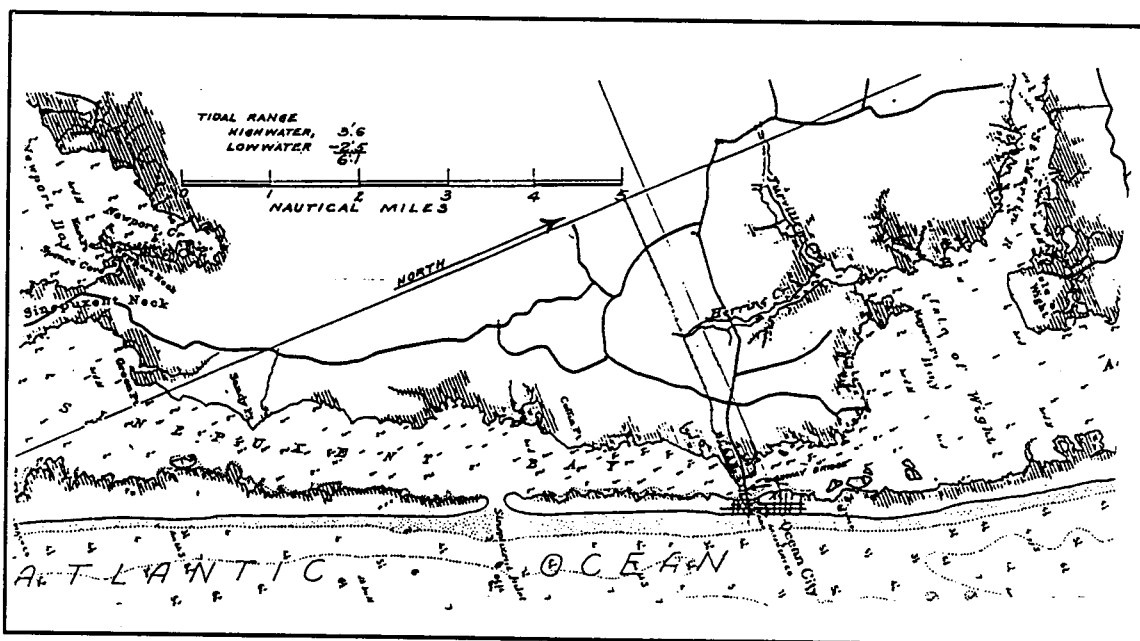


Figure 19. Location of Sinepuxent Inlet in 1920 (Hite 1924).

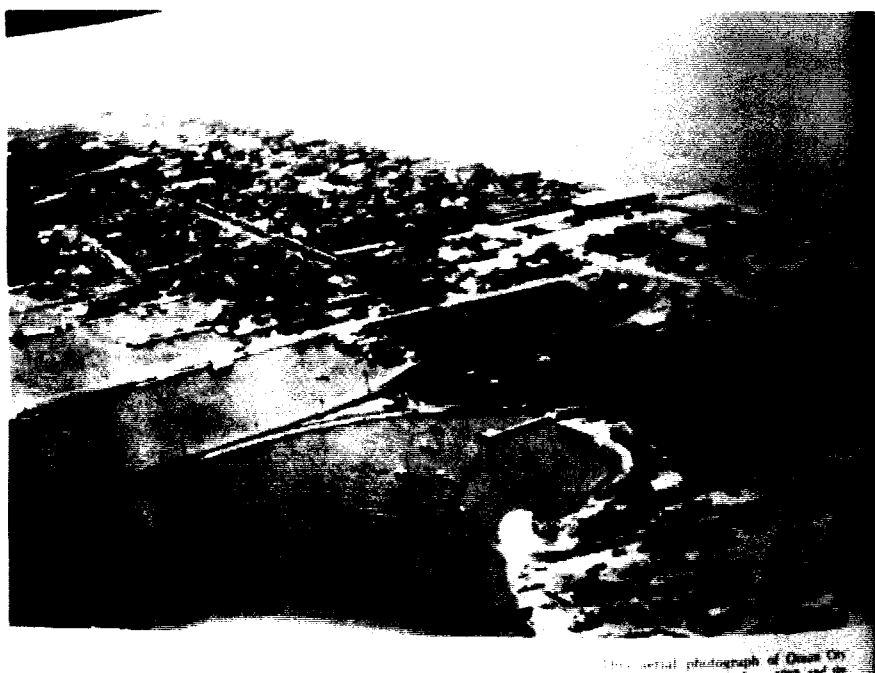


Figure 20. View of approximate location of Ocean City Inlet prior to August 1933 storm. Inlet breached at location of L-shaped dock (Hurley and Hurley 1979).

of a road down the length of the island and for controlled commercial development (National Park Service 1985). In 1976, the National Seashore Act was amended to remove the road and commercial development provisions.

Assateague Island National Seashore extends from Ocean City Inlet, Maryland, to Chincoteague Inlet, Virginia, covering approximately 15,600 hectares of land and water. Within the authorized boundaries of the seashore, three governmental agencies share management jurisdiction: Assateague State Park is managed by the Maryland Park Service. Chincoteague National Wildlife Refuge is managed by the U.S. Fish and Wildlife Service. The rest of Assateague Island National Seashore is managed by the National Park Service (Figure 21). The primary mission of the National Seashore is to provide the public with appropriate outdoor recreational opportunities, and protect and preserve the indigenous natural resources and natural ecosystem conditions and processes. Assateague Island is a strategic location along the Atlantic flyway, the migratory route for various species of birds, including Canadian Geese and Peregrine falcons. The National Seashore is visited by more than 2 million people per year (National Park Service 1985).

Engineering Activities

The boardwalk, various hotels, cottages and beachfront along the adjacent shoreline of Ocean City were heavily damaged after the storm of February 1920. The reconstruction of these cottages, commercial properties, and nearshore beach line initiated the local township's commitment to maintaining their beachfront through various engineering activities. Since 1920, an assortment of engineering practices have been implemented to help combat shoreline erosion for both Fenwick and Assateague Islands. These include; construction of groin fields (both of timber and stone), removing sediment from the foreshore and placing it on the backshore, re-vegetation and sand fence construction along existing dune lines, beach nourishment, and sand tightening of the south jetty at Ocean City Inlet.

From 1922 - 1961, 25 timber groins were constructed at various intervals along the Maryland portion of Fenwick Island. An additional eleven timber groins were constructed from 1972 to present. No quantitative data on these structures was available to support their usefulness on slowing shoreline erosion. Forty-three sand asphalt groins were built during 1954 and 1955. One example of the groins effectiveness to trap sand comes from Shepard and Wanless (1971), they state that during 1954-1955 construction period "so much sand was trapped that many groins were covered within a week of construction". Quantitative information regarding the volume of sediment being trapped by the groins wasn't documented. However, visual accounts give testimony that the groin fields were a very effective littoral barrier. However, the net effect of the entire groin system on the shore has been negligible.

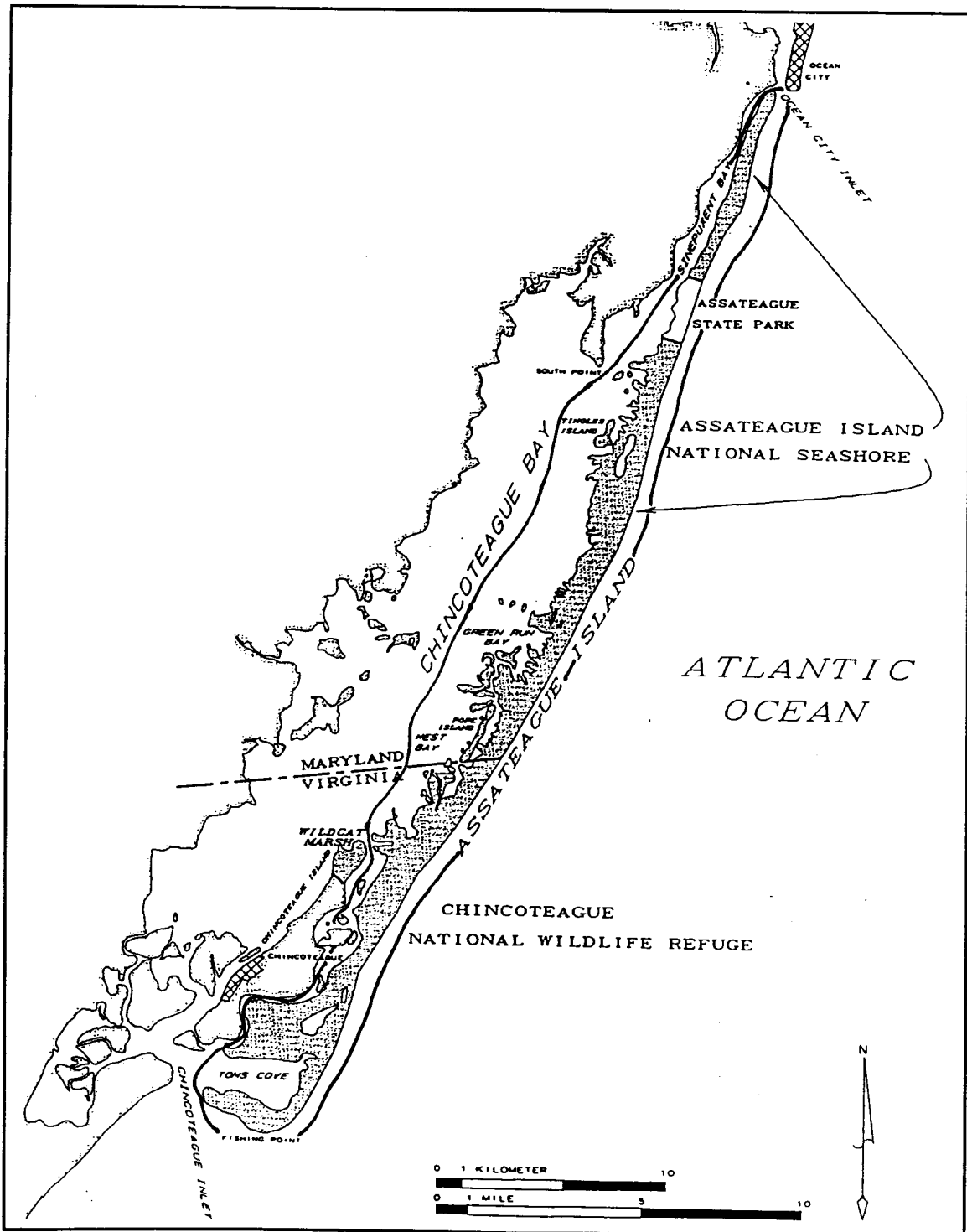


Figure 21. Land and water boundaries surrounding Assateague Island National Seashore.

Heavy machinery, in particular bulldozers, have been used extensively for backshore and foredune development along Fenwick Island. This method was used extensively after the March 1962 storm to reconstruct, heighten and stabilize overwashed dune areas. Beach sand was "borrowed" from the foreshore portions of the beach profile for reconstruction. Beach profiles, taken before and after bulldozing, has shown that the landward-most berms and dunes are building up, at the expense of steepening the foreshore "borrow" area (USACE 1986b).

Fourteen separate dredging events have taken place at the inlet since the inlet opened in 1933 (Table 4). The first dredging event took place in August 1935, along the centerline of the inlet, with an original channel depth of 2.6 m MLW, and channel width of 61 m. The first *maintenance* dredging occurred in 1948, removing approximately 47,880 m³ of sediment. Dredging totals were approximately 5,700 m³ from 1948 to 1963, 11,400 m³ from 1963 to 1970, and increased to 58,520 m³ from 1970 to 1973. After 1973, the annual dredging rate has reduced to approximately 16,720 m³/yr.

Following the March 1962 storm, emergency beach nourishment operations were begun for Ocean City, by the U.S. Army Corps of Engineers, from 2nd Street north to the Maryland-Delaware state line. Approximately 798,000 m³ of sand were deposited along the beach for dune and berm reconstruction, accompanied by sand fencing along the constructed dune ridges. This sediment was borrowed from Isle of Wight and Assawoman Bays. Approximately 760,000 m³ of sediment from channel maintenance dredging was used to close a breach along the south jetty's southwest side. This same location has had approximately 228,000 m³ of sediment placed over the last 10-15 years (USACE 1986b).

The U.S. Army Corps of Engineers, the State of Maryland, Worcester County, and the Town of Ocean City began an extensive beach nourishment project, formally known as "Atlantic Coast of Maryland (Ocean City) Shoreline-Protection Project. This massive project (approximately 11 km in length) was to be constructed in two separate phases. The first phase, initiated by the State of Maryland in 1988, placed a recreational beach fill (2.1 million m³) between 3rd street and the Maryland-Delaware state line (Figure 22). The federal portion (USACE) began in 1990-1991, placing a storm-protection beach fill (2.9 million m³) along this same stretch of shoreline. Other components of this plan included a protective dune (height of 4.42 m NGVD, and 7.6 m wide) extending from 27th Street north to the Delaware State line, vegetative plantings, sediment trapping fences, pedestrian and vehicular crossings, a steel sheetpile bulkhead (from Division St to 27th), and widening of the beach and berm design originally completed by the state of Maryland (Grosskopf and Stauble 1993). Sediment was dredged from 2 linear shoals (shoals 2 and 3) located about 3.2 km east of southern Ocean City, and 4.8 km east of northern Ocean City, respectively (Figure 22). Respective grain sizes were approximately 0.30 to .35 mm, classified as fine to medium sands (Folk 1980). Replenishment activities are scheduled for every four years, assuming an average erosion rate of 101,080 m³ per year (Grosskopf and Stauble 1993).

A comprehensive monitoring plan was established to quantify the long-term behavior of the engineering design and response of the beach fill to large-scale storm events, in particular,

Table 4 Dredging history of Ocean City Inlet from 1935-1990.			
YEAR	CUBIC METERS	AVERAGE ANNUAL	CUMULATIVE
¹ 1935	-	-	-
³ 1947-48	47,880	3,683	47,880
² 1954-55	57,000	8,143	104,880
² 1962-63	28,880	3,610	133,760
² 1969	42,560	7,093	176,320
² 1970	37,240	37,240	213,560
² 1971	98,040	98,040	311,600
² 1973	76,760	38,380	388,360
⁴ 1976	22,800	7,600	411,160
⁴ 1977	34,200	34,200	445,360
⁴ 1978	45,600	60,000	490,960
⁵ 1979	38,000	38,000	528,960
⁵ 1982	26,600	8,866	555,560
⁵ 1990	37,240	4,655	592,800
¹ First dredging event - no dredging quantities available Dredging data compiled from Wicker (1974) ² , Dean et al (1979) ³ , USACE -Baltimore (1982) ⁴ , Jeff Price, Navigation Branch USACE Baltimore District Pers. communication (July 1992) ⁵ .			

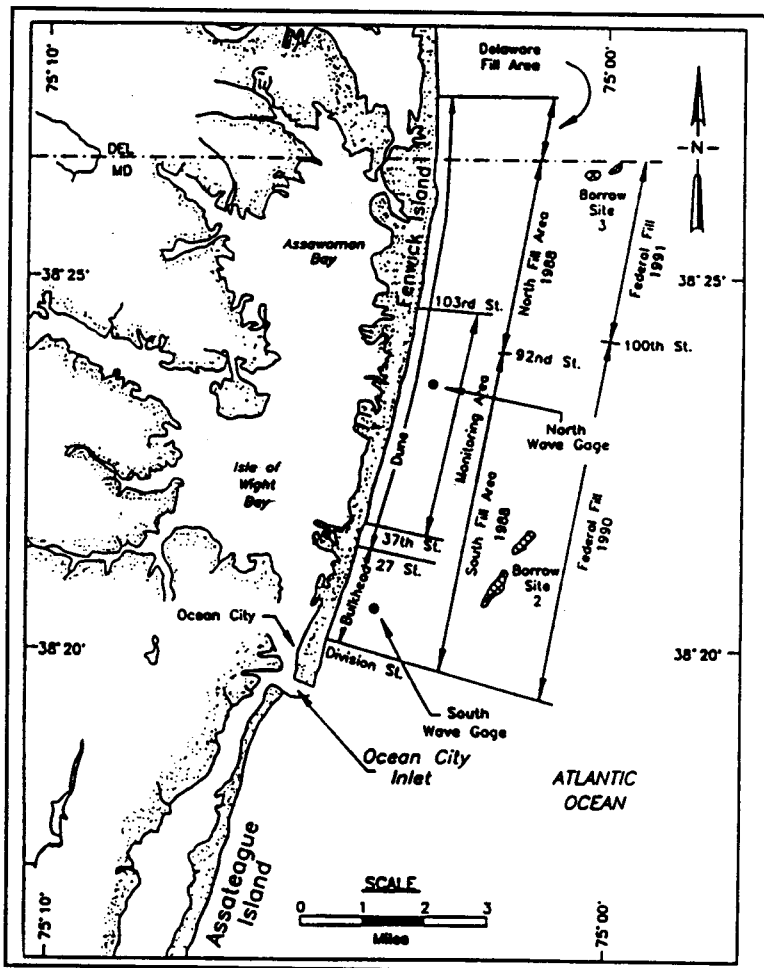


Figure 22. Location of beach nourishment activities along the northern portion of the study area (From Stauble and Kraus 1993).

understanding beach fill evolution under diverse wave conditions, wind and wave currents, and various water level changes. Initial monitoring plans called for profile and sediment samples for 12 sites located in the central portion (5.9 km). Ten additional profile lines were added (totalling 22), with a sampling interval of three times a year, with sediment samples taken along the entire project length. The monitoring plan also called for impromptu surveys after severe storm events (Stauble and Grosskopf 1993). Two major storms impacted this area on October 31, 1991, and January 4, 1992. In general, project performance data revealed removal of portions of the sub-aerial beach (which included the storm dune) offshore within the active profile envelope. Profile surveys conducted in March 1992 showed that substantial quantities of sand had been transported back onto the subaerial beach (Stauble et al 1993). Grosskopf and Stauble (1993) report that "approximately \$93,000,000 in wave, flooding and erosion damages were prevented as a result of the fill project during the two storms. This value does not include nearly \$500,000,000 in revenues and associated taxes from the 1992 Ocean City summer recreational season".

Stabilization of Ocean City Inlet was authorized by the River and Harbor Act of 30 August 1935 and modified by the River and Harbor Act of 3 September 1954. Pertinent project documents are: Rivers and Harbors Committee Document 38, House of Representatives, 72nd Congress, 1st Session; Rivers and Harbors Committee Document 60, House of Representatives, 74th Congress, 1st Session; and House Document 444, 82nd Congress, 2nd Session. The work of improvement authorized by the Act of 30 August 1935 was included in the Public Works Administration Program of 16 September 1933 and the Emergency Relief program of 1935. Construction of the north jetty was initiated in September 1933 and completed by October 1934 (Dean et al. 1979). South jetty construction began in October 1934, and was completed eight months later in May 1935. Since 1938, structural modifications have been necessary due to sediment buildup/impoundment, and breaching of the inshore section, for the north and south jetty, respectively. Figure 23 gives overall project dimensions and chronology of construction events for the north and south jetties from 1933-present. In December 1984, rehabilitation began on the south jetty to "sand tighten" the porous rubble mound material used during construction. This would impound sediment, thereby reducing the danger of breaching the inshore portion of the south jetty tie-out. Wave refraction around the ebb tidal delta reversed littoral drift movement along the shore, moving sediment north towards the inlet throat. Wave overtopping, structure permeability, and low crest elevation of the south jetty were responsible for allowing sediment to be transported through the structure. Consequently, shoaling problems were occurring within the navigation channel along the northwest portion of Assateague Island. In order to eliminate or significantly reduce the problem shoal, construction began on a new jetty section along the southern side of the old existing structure. The rehabilitation consisted of layers of bedding material, corestone, one layer of capstone, and precast concrete units along the centerline of the structure (Bass et. al., 1994). After jetty rehabilitation was complete, concerns over erosion of the northwestern end of northern Assateague arose due to elimination of this source of sand. To stabilize the northern end of Assateague Island, three headland breakwaters were constructed (Figure 23). Breakwater no. 1 was 103.6 m in length and was tied into the existing south jetty. Breakwaters no. 2 and 3 were spaced 91.4 m apart from Breakwater 1, and were each 60.9 m long. The south jetty rehabilitation project was completed in December 1985. Prior to the sand

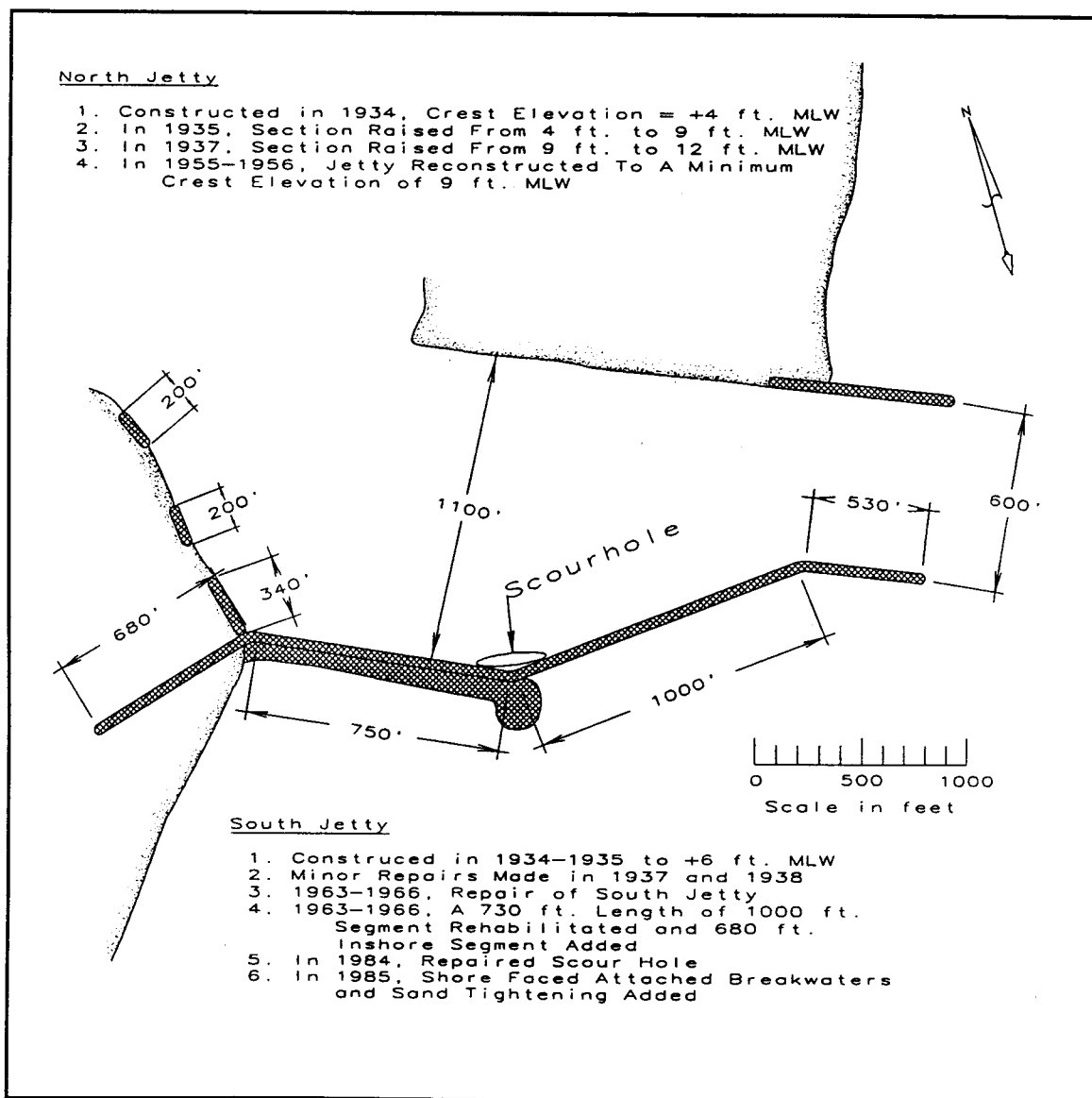


Figure 23. Chronology of construction events for Ocean City Inlet from 1933-present (Modified from Dean et al 1979).

tightening project, south jetty rehabilitation also included filling an existing scour hole along the inner portion of the south jetty. The channel thalweg (centerline of channel) had migrated towards the south jetty since stabilization took place in 1935. The scour hole had deepened to approximately 15.2 m by 1984. Figure 24 shows comparisons of a bathymetric survey completed in 1984 (immediately after project completion) and 1990. These comparisons confirm that no further scouring has occurred since project completion in 1984.

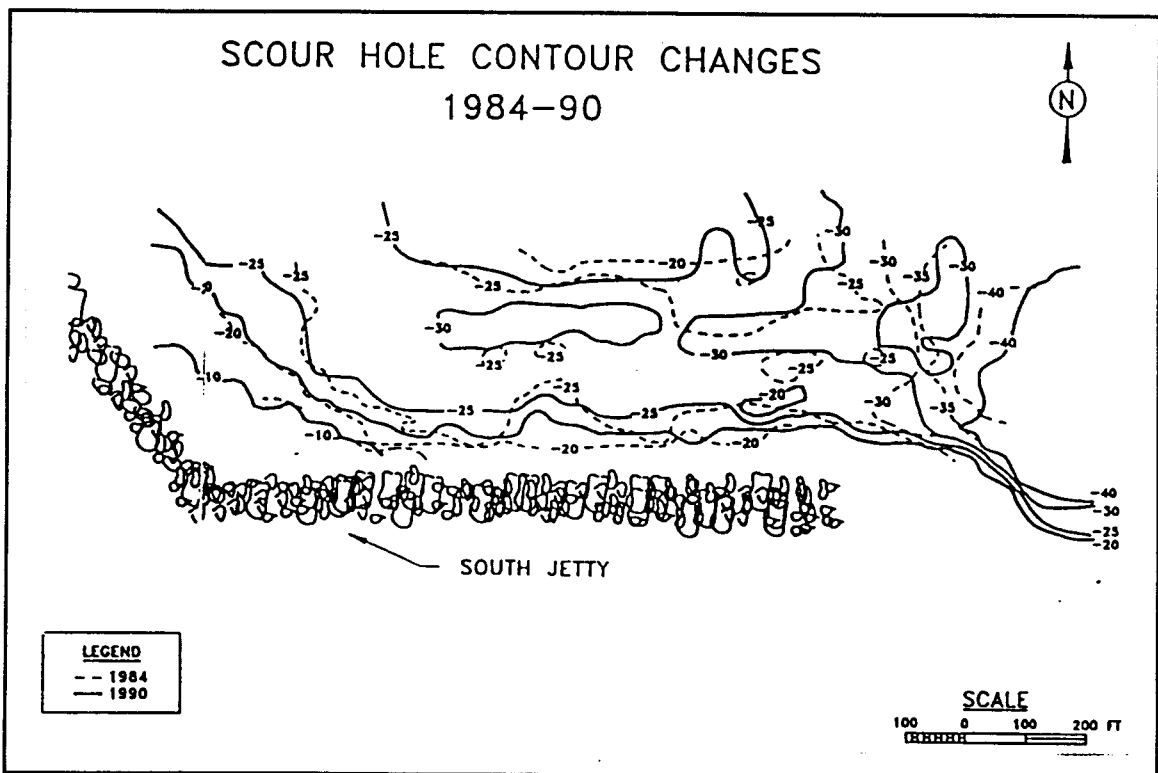


Figure 24. Location of scour hole adjacent to the south jetty (From Bass et al 1994).

4 Inlet Morphodynamics

The term "morphodynamics" is derived from morph, meaning form, shape, and/or structure, and the term "dynamics", which is the study of the relationship between motion and the forces affecting motion. Quantifying an inlet's "morphodynamic" behavior requires an understanding of its origin, wave and tidal relationship, associated sand bodies seaward and landward of the inlet, tidal prism and inlet cross sectional area (stability characteristics), and influences of human activity surrounding the inlet. Tidal inlets are complex, dynamic environments influenced by sedimentary processes, which tend to close and/or cause lateral migration, and hydraulic processes which tend to keep the inlet open. Advancements in cartographic and hydrographic survey techniques, and utilization of state-of-the-art Geographic Information Systems (GIS) for quantifying shoreline and bathymetric changes, has provided coastal engineers and scientists unprecedented capabilities to quantitatively document morphodynamic changes associated with tidal inlets.

An exhaustive quantity of scientific literature pertaining to tidal inlets has been published (for extensive bibliographic references see Barwis 1976, USACE 1984, Weishar and Fields 1985, Aubrey and Weishar 1988). The vast quantities of data generated from these studies and others, have revealed global similarities in tidal inlet development, morphology, and sediment patterns (Hayes 1980).

A key element in understanding how the development of Ocean City Inlet ebb shoal has affected Northern Assateague Island, is the integration of hydrodynamic and sediment dynamic information with various engineering activities since the inlet opened in 1933. The following sections will review and highlight coastal processes and engineering activities that relate *specifically* to the Ocean City Inlet study area.

Inlet Origins

Natural tidal inlet formation is directly linked to high energy events such as tropical or extratropical storms (Johnson 1919, Hite 1924, El-Ashry and Wanless 1965, Pierce 1969). Inlets are often created by elevated storm surges and above normal wave forces. Physical factors such as barrier width, barrier height, magnitude and duration of storm, depth of lagoon adjacent to the barrier, size of lagoon, and number of current openings from lagoon to ocean, will determine the vulnerability of a barrier island to inlet creation. Inlet formation by barrier breaching can occur from either the ocean side or lagoon side. Ocean side inlet formation can result from focusing direct frontal wave attack along a very narrow stretch of shoreline. Overtopping of the storm beach ridge by elevated storm surges causes a breach in the barrier that creates a connection between the ocean and lagoon adjacent to the barrier. Inlet formation from the lagoon side usually occurs by breaching from the backside of the barrier island. Elevated storm surges, increased by strong wind and wave energies, move water onshore across the barrier, elevating water levels in

the back lagoon. If the storm moves rapidly north, its counterclockwise wind circulation can abruptly reverse winds from onshore to offshore. The abnormally high mass of water in the lagoon is forced seaward across low areas of the barrier, scouring a channel and connecting the lagoon with the ocean (Figure 25). Typically, a combination of these factors results in new inlet formations. For example, frontal wave attack and elevated storm surges overwash and lower the height of the storm beach ridge. Cutting of a new channel from the lagoon to the ocean would be facilitated through previous overwash channels, and prior erosion of the front of the barrier. The process of inlet breaching, migration, closing and development of a large flood tidal delta, as shown in Figure 26, is typical of inlet processes at Assateague Island. The process of inlet formation from the lagoon side (ebb surge breaching) has been cited as the method by which Ocean City Inlet was created. Truitt (1968) attributed Ocean City inlet's formation to "the outflow of pent-up high water in Assawoman and Isle of Wight Bays...". Another example of natural inlet formation within the study area is given by Hite (1924). He describes the formation of Sinepuxent Inlet (located approximately 5 km south of Ocean City Inlet) during a major northeaster in 1920. This sequence of events is depicted in Figure 27. His first observations center around strong wind and wave conditions, and severe overwashing of the island. In his words, he describes "the beach was afloat, literally, and the sand alive". Later, the wind began to shift from north-northeast to north-northwest. The rain and winds slowed, forcing lagoonal water back over the barrier island "like the tilting of a pan, a great portion of the waters of the lagoon slid off into the sea and the beach emerged as separate embryo inlets". Two days after the storm, the lagoon waters had lowered through the numerous channels connecting the lagoon to the ocean. By the end of the third day only two channels remained open, less than 8 km apart from each other. Eventually, the southern most inlet closed, leaving Sinepuxent inlet open. This inlet remained active until 1928, when it was closed by another northeaster.

Based on the last 100 years, the 86 km shoreline from Cape Henlopen Delaware to Fishing Point Virginia, had the highest density of historical and active inlets (0.29 inlets/km) in the U. S. (McBride and Moslow 1991). A number of factors contribute directly to tidal inlet frequency, duration, and size (Figure 28). These variables will be described individually in the following sections. Cumulatively, these physical elements affect the degree to which inlet processes alter barrier island shorelines (FitzGerald 1988). Johnson (1919) believed that tidal inlet formation and duration is directly related to proximity to their source of sediment. He assumed that the closer the source of sediment, the less likely any newly created inlet would stay open, due to infilling/deposition from longshore drift. His assumptions also included greater likelihood of lagoonal infilling where there were numerous inlets by providing a uniform salinity for establishment of back barrier vegetation. The proximity of Ocean City Inlet to its source of sediment (i.e. Bethany Beach headland, approximately 16 km to the north) required jetty construction to prevent closure by longshore drift.

Truitt (1967), and McBride (1986), have documented and referenced various locations of temporary inlets along Fenwick and Assateague Islands (Figure 29). At least 11 inlets have been cut through Assateague Island since the early 1800's, eight within its present boundaries (Truitt 1967). These inlets tended to be ephemeral in nature, originating during storm conditions and

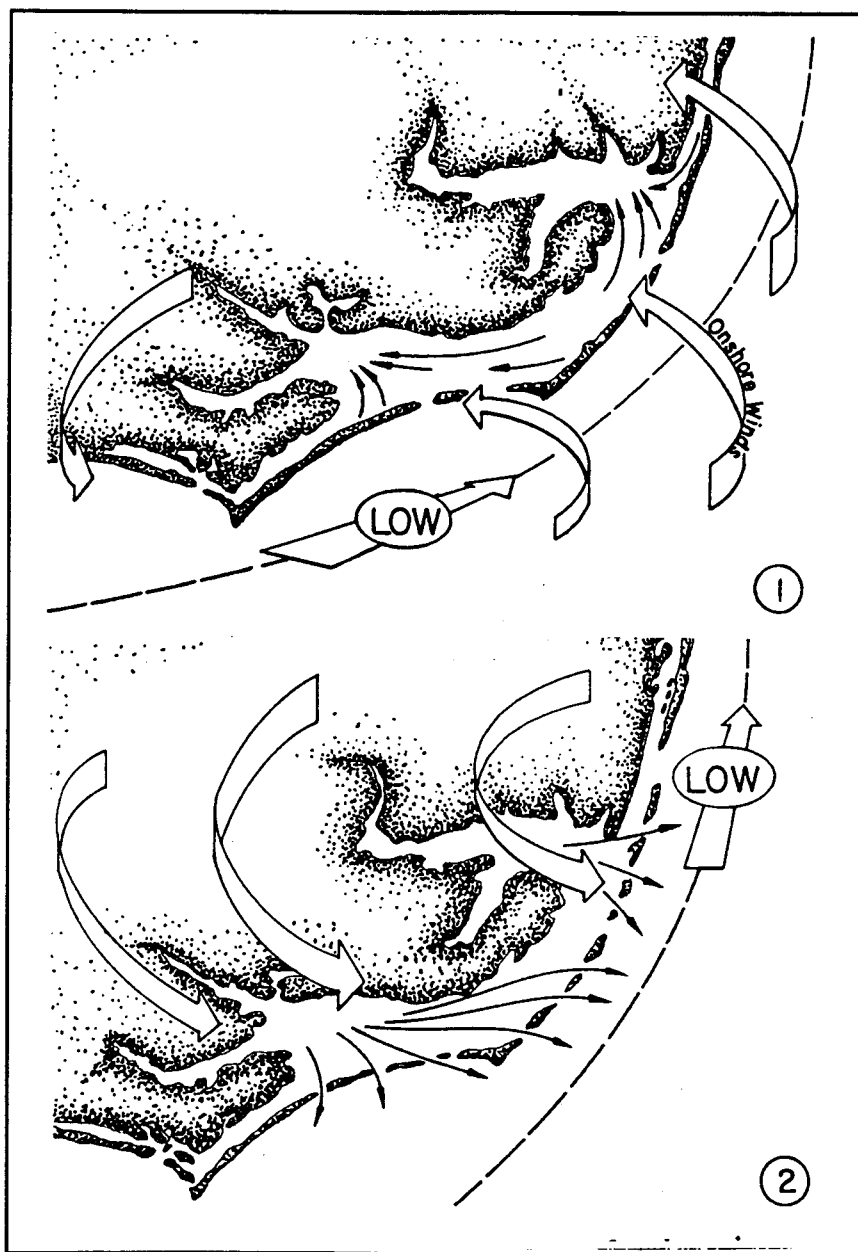


Figure 25. Creation of an inlet by bayside breaching following the passage of a storm event. Note shifting of winds from onshore to offshore as low pressure center passes along the coast (From Leatherman 1988).

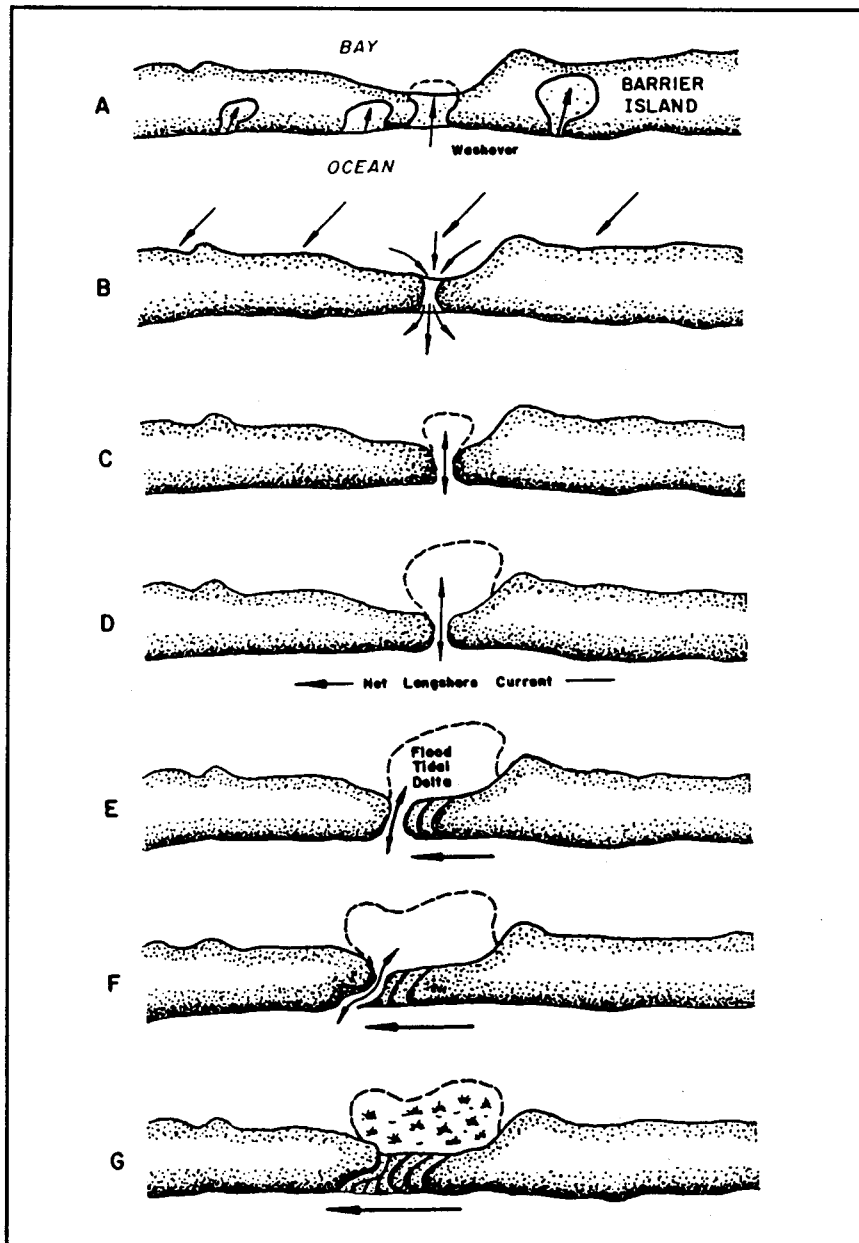
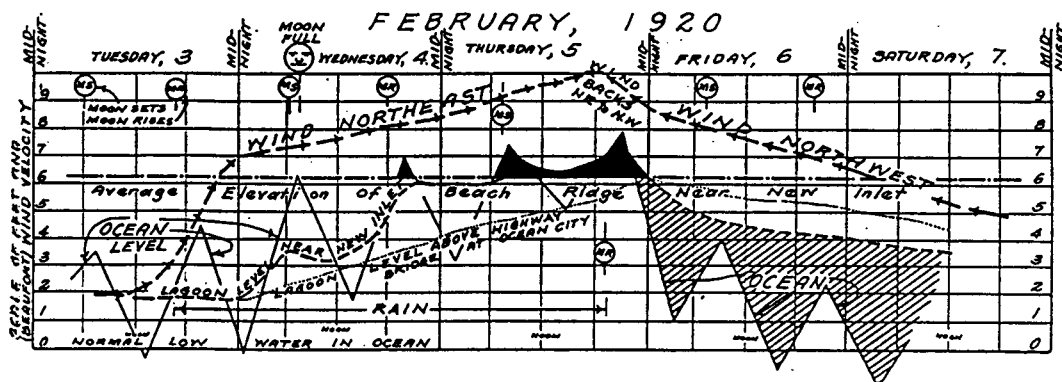


Figure 26. Sequential diagrams illustrating the processes of inlet formation and development of a flood tidal delta, typical of events at Assateague Island (From Leatherman 1988).



THE BIRTH OF SINEPUXENT INLET, MARYLAND.

The initial storm winds slowly banked the ocean water against the beach until the comparatively small lagoon was flooded. When the wind changed a veritable flood was hurled against the beach ridge and numbers of new inlets appeared. Two of these stabilized and one, Sinepuxent, remains to this day. Solid black represents water elevated above the beach ridge and available for cutting new inlets. Shaded areas represent water in the lagoon above the ocean level and available for eroding inlet channels after they have been started. Note the peculiar action of the highway bridge at Ocean City, Md.

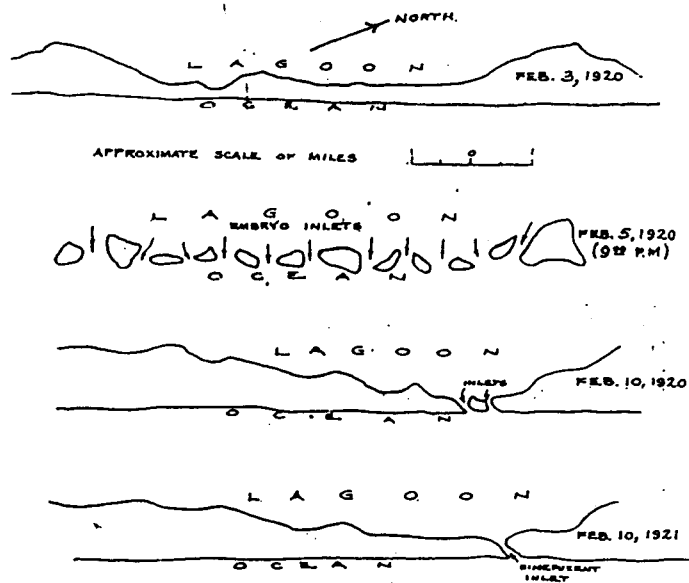


Figure 27. Testimony and field notes from Hite 1924, on the creation of Sinepuxent Inlet after severe northeaster in February 1920.

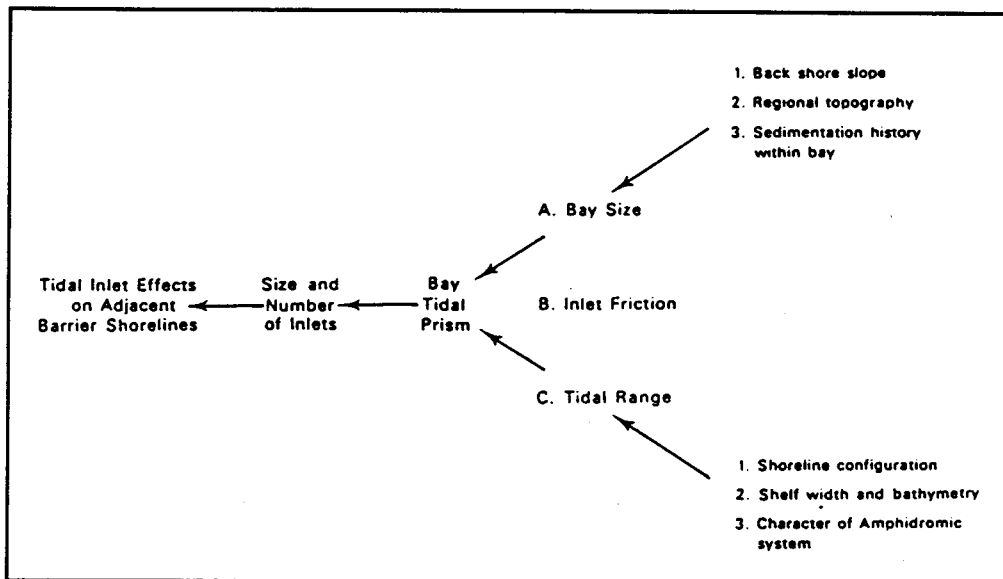


Figure 28. Contributing factors associated with inlet frequency, duration, and size (From FitzGerald 1988).

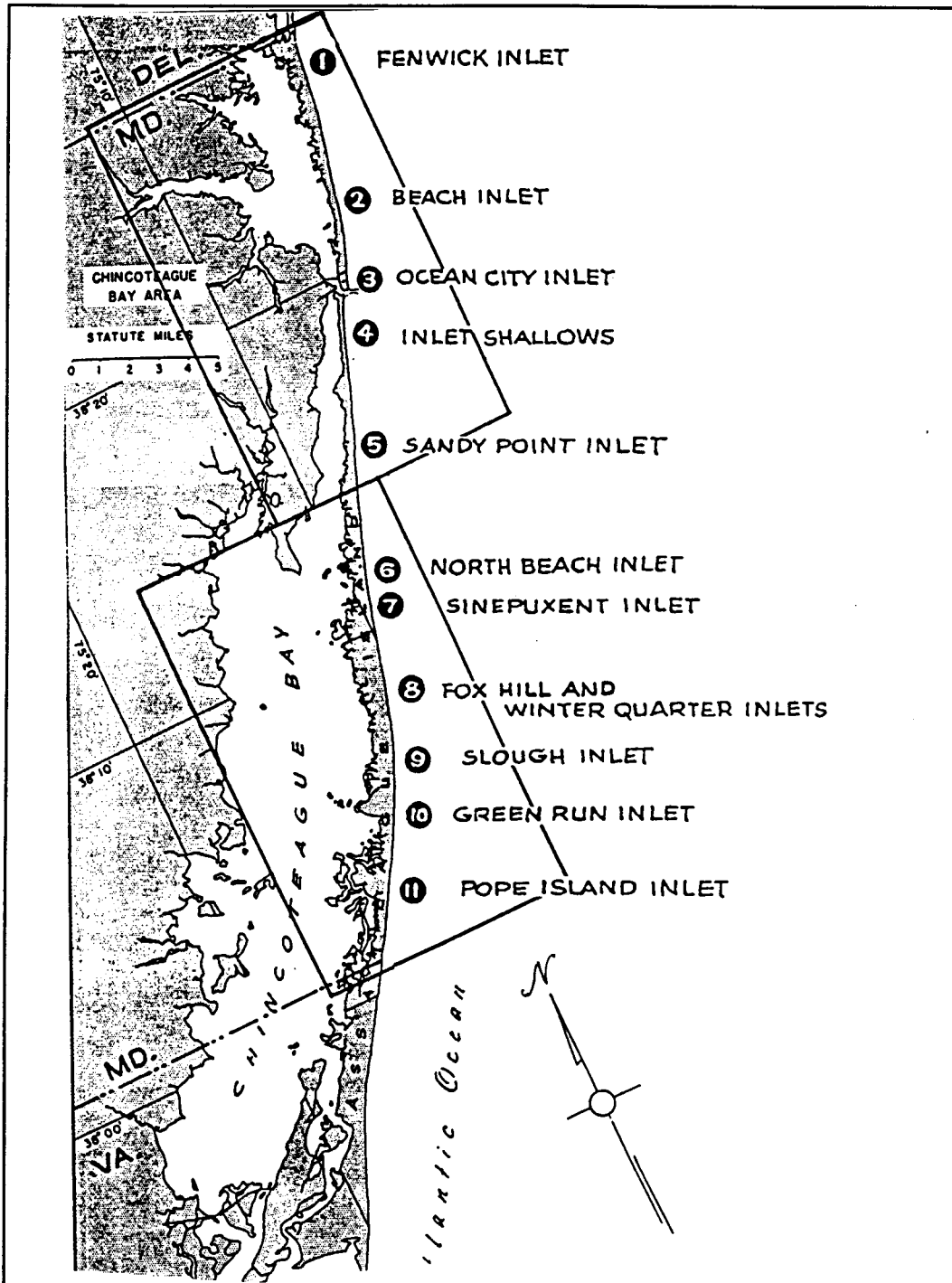


Figure 29. Various historical inlet locations along Fenwick and Assateague Islands (From Truitt 1967).

eventually closing due to a combination of natural coastal processes (i.e. incident wave conditions and longshore drift). Within three miles of Ocean City Inlet's present location were Beach Inlet and Inlet Shallows. Beach Inlet was located approximately three miles north of Ocean City Inlet in the vicinity of a relic paleochannel of the ancestral St. Martin River (Toscano et al, 1989) (Figure 30). Inlet Shallows is a descriptive term for a 1.2 km wide area south of Ocean City Inlet that is notorious for becoming overwashed and/or breached during storm conditions. Prior to the severe storm of 1933, Congress had authorized construction and stabilization of an inlet, just below old Sinepuxent Inlet (approximately 8 km south of present inlet). Then, on August 23, 1933, Ocean City Inlet was created by ebb surge breaching of Fenwick Island spit, scouring the bottom to a depth that was almost immediately navigable to commercial vessels (approximately 3 m deep and 76 m wide). After reviewing the situation, Congress decided to stabilize the new inlet at its present location. Examination of the new inlet's location revealed that the breach occurred adjacent to an existing timber groin. Construction began on the north jetty by depositing stone on both sides of this structure.

Wave and Tidal Relationships

Tidal inlet sand bodies may comprise 30-60 percent of littoral material deposited in a barrier island complex (Hayes 1980, FitzGerald 1988). The accumulation and distribution of these sediments, either landward or seaward of inlets, are primarily controlled by the interaction and/or dominance of wave and tidal energy. Periodic ebb and flood reversals, sedimentation patterns, and incident wave climate, are all part of a self-adjusting control system producing "signature" tidal inlet morphologies, depending on the dominant force. Price (1955), was the first to establish a relationship between wave dominated versus tide dominated coastlines. Other publications followed, notably Davies (1964), Hayes (1975 and 1979), and Davis and Hayes (1984), further elaborating on this form-process relationship. Collectively, this information was used to develop a common classification scheme for describing the dominant force acting upon the various shorelines. This information was further used for reconstruction of stratigraphic sequences in the ancient record for depositional history information, provides present scientists and engineers morphological "imprints" of dominant forces acting along a study area, and allows a global perspective for understanding and comparing similarities of coastal environments.

Tidal range has been classified as either Microtidal (0-2 m), Mesotidal (2-4 m), or Macrotidal (> 4 m) (Hayes 1975). Figure 31 shows the relationship of tidal range to coastal morphological features. Note that barrier island coasts develop best under microtidal conditions, tidal deltas and inlets in mesotidal conditions, and tidal flats and salt marshes in areas having macrotidal tidal ranges. Microtidal shorelines are characteristically long, straight and narrow, have few tidal inlets, and display numerous washover features. Typically, these barrier islands become overwashed during storm events. Rising storm surge waters are unable to flow past the barriers due to the absence of tidal inlets. Therefore, topographic lows across the barrier become conduits for the storm surge, resulting in multiple washover fans and/or ephemeral inlets (Hayes 1979). Mesotidal

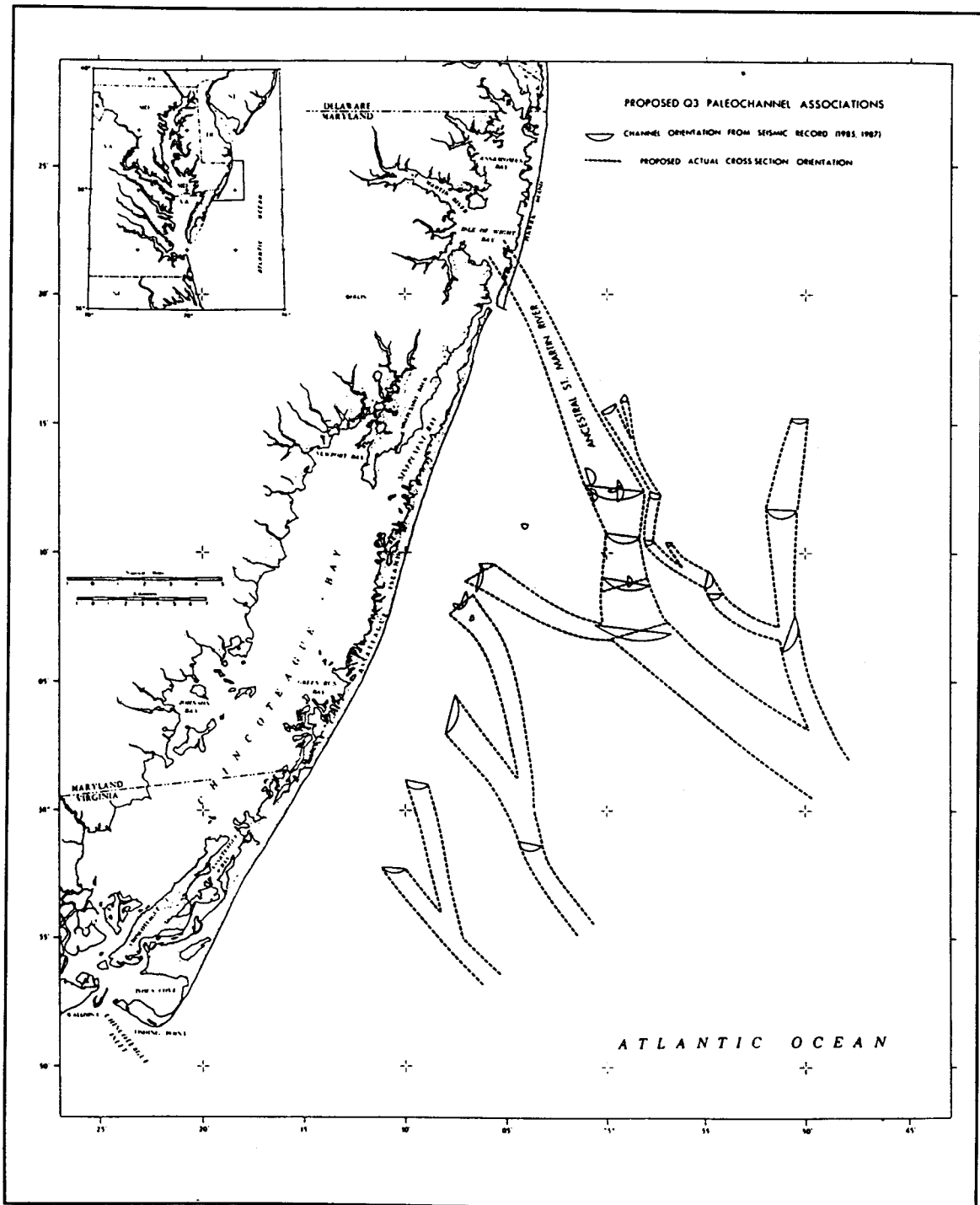


Figure 30. Location of relic ancestral St. Martin River (From Toscano et al 1989).

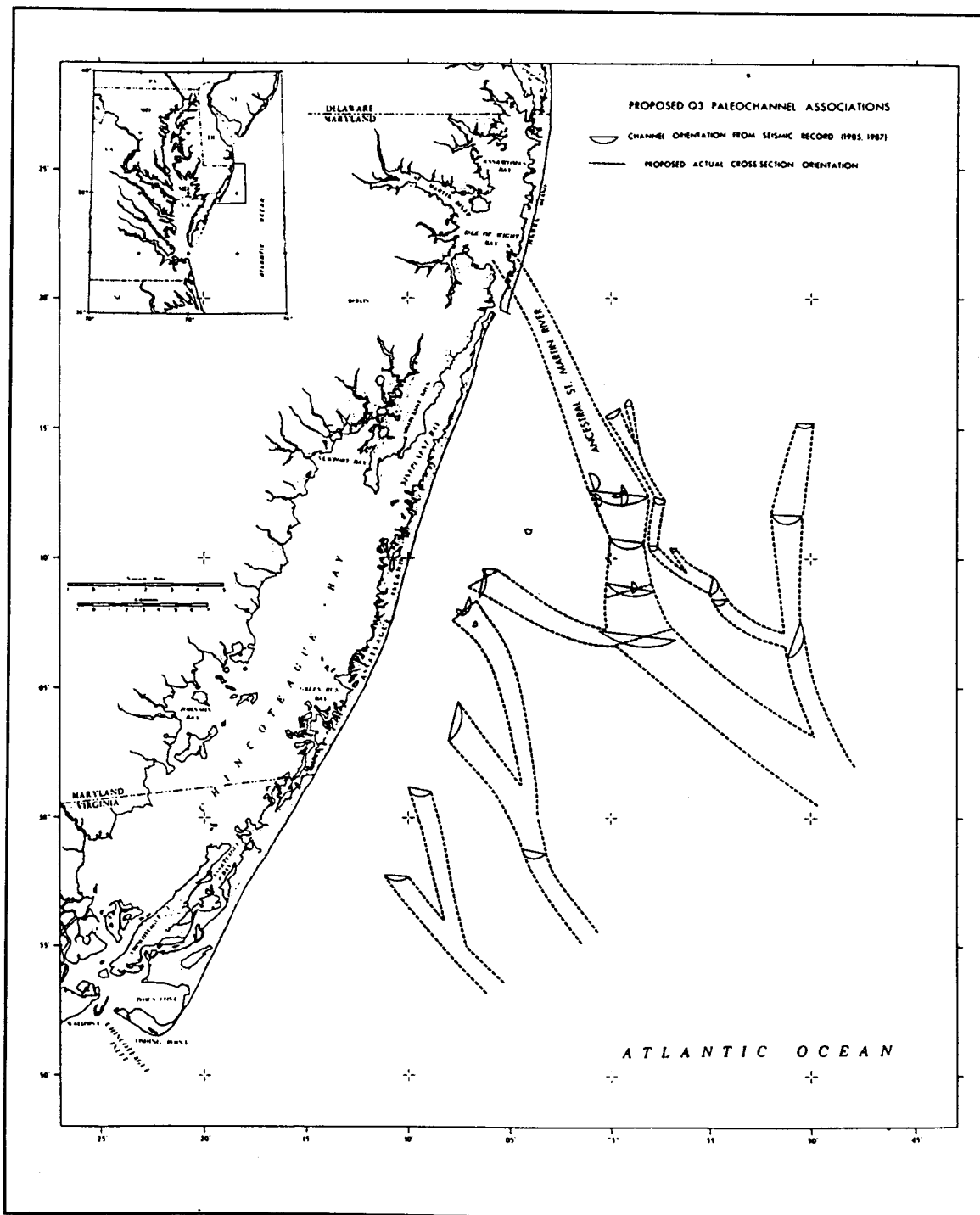


Figure 30. Location of relic ancestral St. Martin River (From Toscano et al 1989).

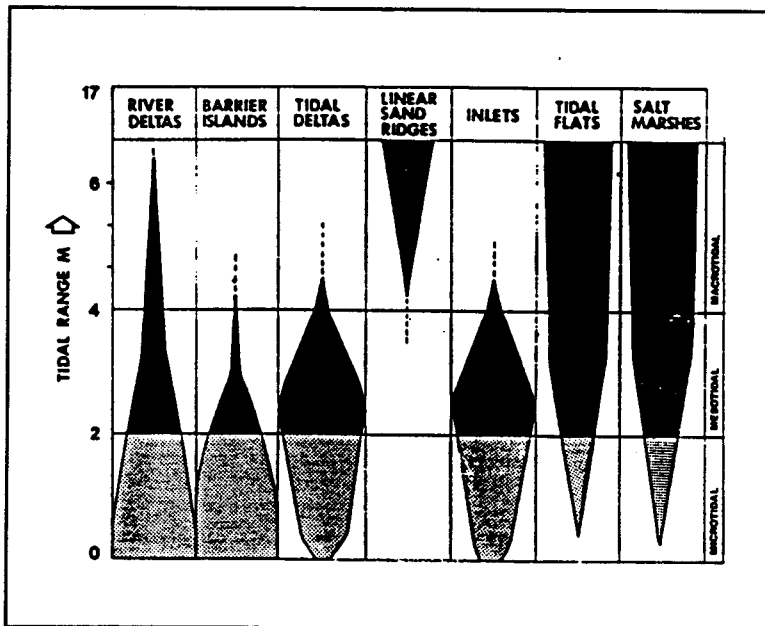


Figure 31. Variation in morphology of coastal-plain shorelines with respect to differences in tidal range (After Davies, From Hayes 1975).

shorelines are short and stubby, may exhibit a "drumstick" configuration, have multiple inlets, and very well developed ebb deltas (Hayes 1979). Macrotidal shorelines lack barrier islands and exhibit long, linear sand bodies within their central and offshore portions. Only 10 % of the worlds barrier islands occur within this tidal range (Glaeser 1978).

Coastal shorelines are also classified as being Wave Dominated or Tide Dominated (Price 1955, Hayes 1975, Davis and Hayes 1984). Overall, wave and tidal energy, when linked directly to tidal range, are the two most critical factors controlling a shorelines morphology. Hayes (1979) states "in areas of average marine wave conditions, coasts with small tidal ranges (microtidal) are usually dominated by wave energy, and coasts with large tidal ranges (macrotidal) are usually dominated by tidal currents and tidal fluctuations. Coasts with intermediate tides (mesotidal) show influences of both waves and tides and are thus termed mixed energy coasts". These general shoreline characterizations are provided as the general rule. That is, exceptions to these descriptions do occur, and have been cited in the literature (Davis and Hayes 1984). One such example is the large number of inlets located on the western Florida coast in the Gulf of Mexico. Wave energy is low with a mean annual wave height of 30 cm. However, there are abundant tidal inlets, large ebb deltas, and characteristic drumstick-shaped barriers. These features are usually associated with mesotidal coastlines. One variable which may account for these exceptions is tidal prism. This is defined as the total amount of water (corresponding to spring range of tide) that flows into an and out of an inlet with movement of the tide, excluding freshwater flow (USACE 1984, Jarrett 1976). Inlet tidal prism volumes may vary along the same tidal range classified stretch of shoreline (micro-, meso-, or macrotidal). An increased tidal prism volume would amplify both ebb and flood inlet current velocities, allowing for greater sediment transport capabilities, as compared to an adjacent inlet with a smaller tidal prism. Figure 32 shows that no relationship exists for tidal prism and tidal range. An interesting note from this figure is that inlets found along the Gulf of Mexico (G) show the highest and lowest tidal prism values, even though tidal ranges are less than 1 m. The lack of correlation between these two variables would explain why well developed ebb-tidal shoals and other common mesotidal inlet characteristics can be found on microtidal or low wave energy coasts (Davis and Hayes 1984).

The Delmarva Peninsula falls within the tidal range of microtidal. The northern and southern portions are classified as mixed energy wave dominated, and mixed energy tide dominated shorelines, respectively (Hayes 1979, McBride and Moslow 1991). The coastal morphology of the northern Delmarva Peninsula (Cape Henlopen, De to Fishing Point, Va) displays features indicative of a microtidal, mixed-energy wave dominated shoreline. These are long, straight, narrow barriers, few tidal inlets, and numerous washover features. A large tidal prism exists for Ocean City Inlet, producing a sizable ebb-tidal shoal that extends approximately 1 km offshore. The southern portions of the Delmarva Peninsula (Chincoteague, Va to Cape Charles, Va) exhibit relatively stable inlets, extensive backbarrier salt marsh complexes (shallow bays, tidal creeks, and tidal flats), and extensive sand bodies seaward of the inlets. These characteristics indicate a more tidally influenced area.

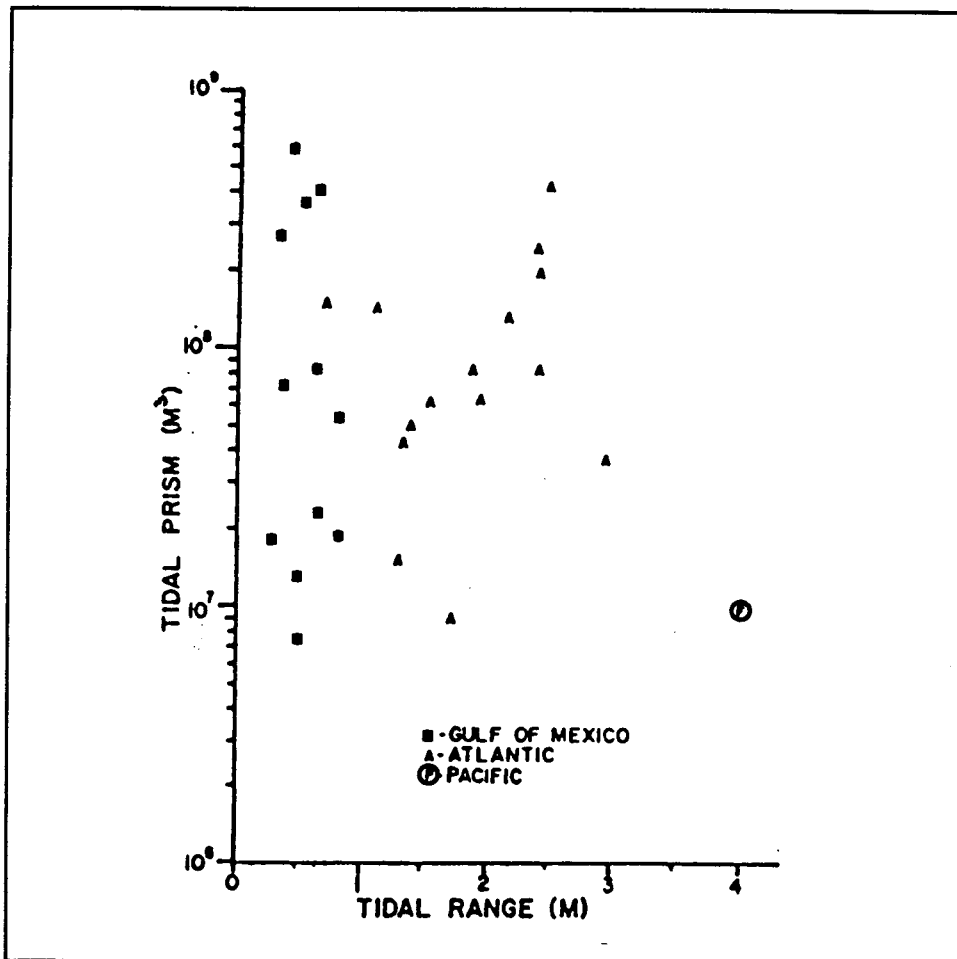


Figure 32. Plot of tidal range and tidal prism for Atlantic and Gulf coast inlets (From Davis and Hayes 1984).

Ebb Shoal Morphology

The terminology used in the previous section has been used to describe an inlets morphology, that is, the absence, presence and location of their associated sand bodies. Hubbard et al. 1979, states "the distinction between inlet types (i.e. wave dominated versus tide dominated) is based on inlet morphology and on the distribution of sand bodies within the system. If the ebb-tidal delta is a result of the interaction between tidal currents and wave energy, then the resulting sand body should reflect the relative importance of each process..". Inlet types were divided into three categories: wave dominated, tide dominated, and transitional (Hubbard et al. 1979). Wave dominated inlets have small ebb tidal deltas, and large multi-lobate flood deltas due to the predominant landward sediment transport. Tide dominated inlets have well-developed ebb deltas, and deeply incised ebb channels flanked by extensive channel-margin bars. Transitional inlets have their sand-bodies confined mainly within the inlet throat region (Hayes 1991). Coastal current (longshore, onshore and offshore) and wave interaction with tidal inlet currents influence the depositional patterns of inlet sand-bodies (Hales and Herbich 1973, Oertel 1975). Figure 33 illustrates four possible ebb delta configurations resulting from these coastal current interactions. The formation of ebb-tidal delta sandbodies prevents littoral drift material from reaching the adjacent shoreline by acting as sediment sinks or traps, accumulating large reservoirs of sand (Dean and Walton 1975, Walton and Adams 1976). Sediment volumes found in these ebb shoals is very significant when evaluated in terms of erosion or deposition of adjacent barrier islands. Walton and Adams (1976) examined 44 U.S. inlets (examples from East coast, West coast, and Gulf of Mexico) for correlations between the amount of sand stored in the outer bar/shoal to an inlets tidal prism, cross sectional inlet throat area, and wave energy exerted on these shoals. A strong trend (2 orders of magnitude) existed for increasing outer bar/shoal storage capacity with increasing tidal prism and cross sectional inlet throat area (i.e. larger tidal prisms promote greater ebb shoal deposition). Inlets with lower incident wave energy had larger outer bar volumes than inlets found on high energy coasts. One explanation is that more available wave energy exists along high energy coasts to force these shoals landward after initially being deposited as a shoal. The tendency of sand to become trapped on outer bar/shoals may depend on other variables such as size distribution of outer bar/shoal sediments, and longshore energy flux (Walton and Adams 1976). The outer bar/shoal system is constantly evolving with longshore sediments providing material to the inlet, ebb currents moving material offshore, and wave energy re-shaping and shifting material towards the inlet. This succession of dynamic activity naturally winnows finer grained sediment through and/or away from the inlet, leaving a coarser grain sediment residing within the outer bar/shoal complex (Postma 1967).

Hayes (1991) presented a model for "typical" morphologic characteristics of an ebb-tidal delta (Figure 34). The primary components of this ebb-tidal delta are: (1) a main ebb channel; (2) channel margin linear bars, built by interaction of flood/ebb tides and wave induced currents, (3) a relatively steep, seaward-sloping terminal lobe located along outer edge of main ebb channel; (4) swash platform, broad sand sheets deposited over the ebb platform that flanks the ebb channel; (5) swash bars, isolated sand bodies, built by swash action of waves that migrate across the swash

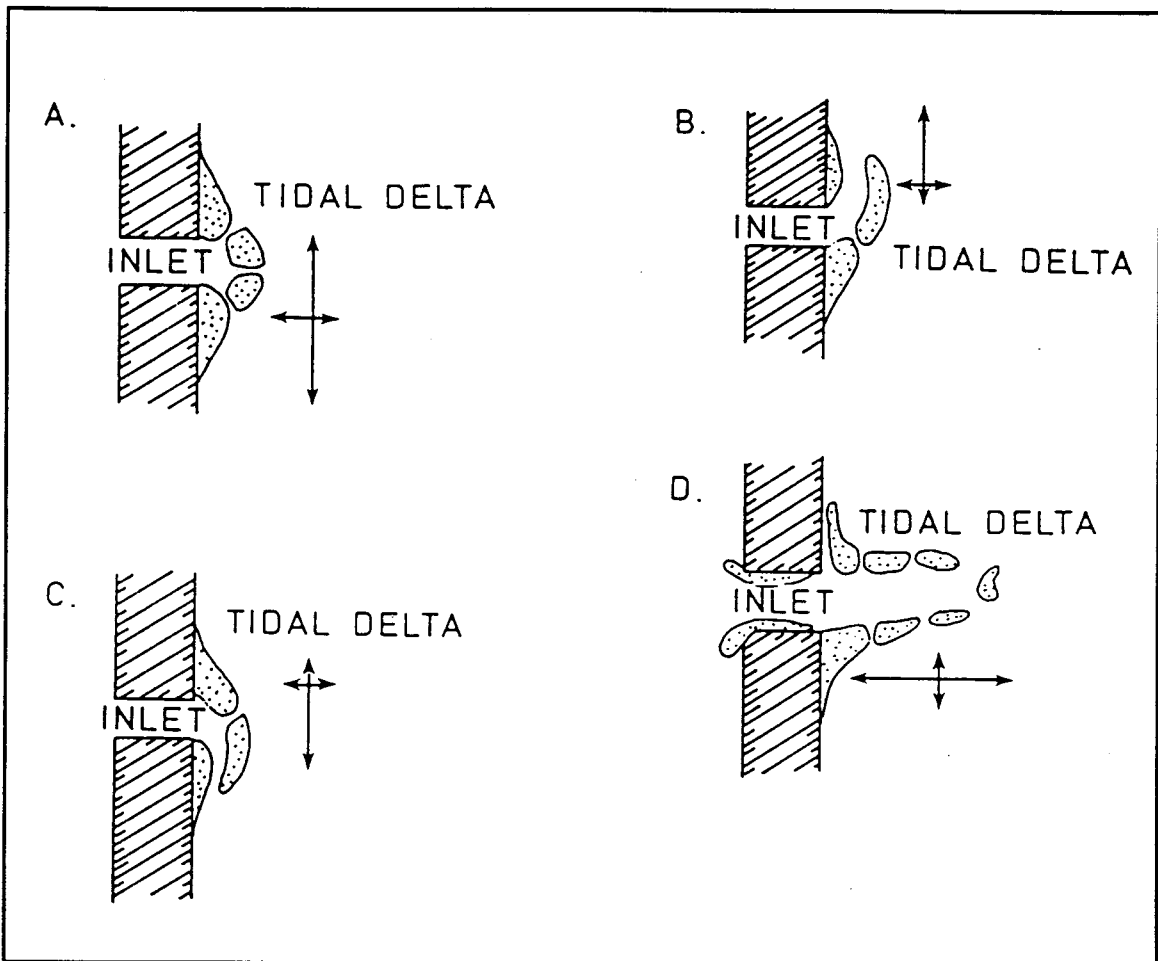


Figure 33. Four possible ebb shoal configurations resulting from wave and current interactions. Arrow signifies prevailing force (longshore, onshore, offshore) (From Oertel 1975).

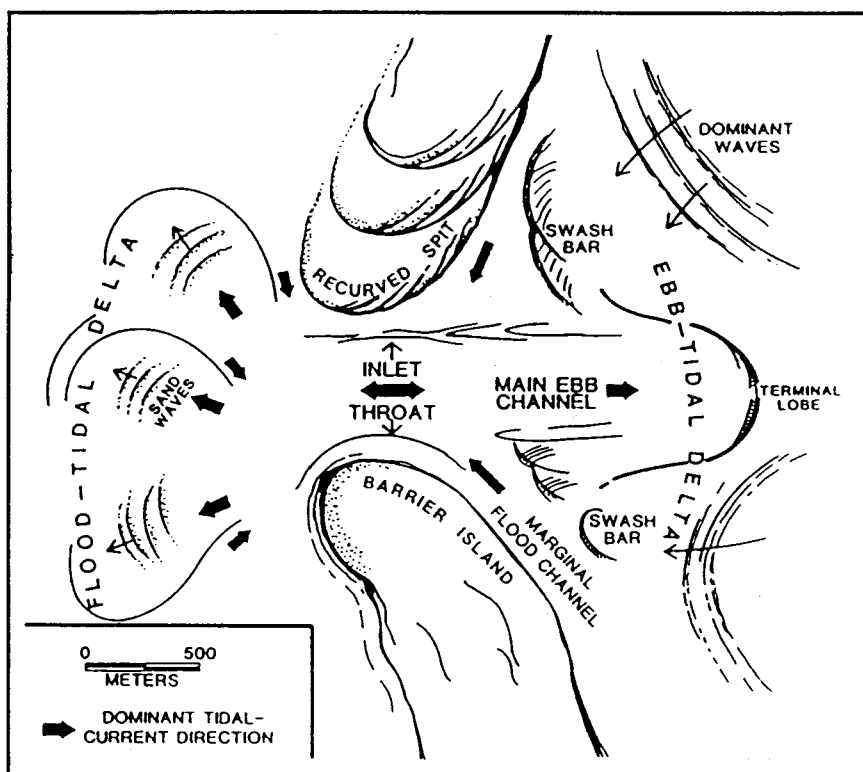


Figure 34. Ebb tidal delta morphology (From Hayes 1991).

platform; and (6) marginal flood channels, which occur between swash platforms and adjacent updrift and downdrift beaches. The morphology of these seaward shoals can also be influenced by a phenomenon called time-velocity asymmetry. As first described by Postma 1967, this phenomenon points out that maximum high and low tidal velocities do not occur at mid-tide (Figure 35). The time-velocity asymmetry of tidal currents into and out of an inlet help sustain the ebb tidal delta morphology over time. Hayes (1980) gives this explanation: "...at low water, as the tide turns, strong currents are still flowing out of the main ebb channel. As the water level rises, the flood currents seek the paths of least resistance around the margin of the delta. This creates the horizontal segregation of flood and ebb currents in the tidal channel that creates or helps shape an ebb tidal deltas morphology".

Since the ebb shoal is typically a shallower area, it serves as a dissipater of wave energy. If the ebb shoal causes waves to break offshore, the adjacent downdrift shoreline is sheltered from the erosional impacts of these waves (Hayes 1977, Reynolds 1985). This "shielding benefit" is particularly important during northeast storm events. Figure 36 shows Ocean City Inlet's ebb shoal dissipating 2 m waves during a northeaster. By causing these waves to break offshore from northern Assateague Island, subsequent beach erosion is reduced. Humphries (1977), conducted a seasonal study of inlet-beach sediment exchanges at North Inlet, South Carolina. Between June 1975 and May 1976, North Island beaches adjacent to the inlet did not suffer severe storm damage. The sheltering by the ebb shoal kept northeast waves from directly striking the coast. In fact, downdrift accretion was shown to occur in the vicinity of the attachment point of the ebb shoal to the beachface. Wave refraction around the inlet ebb-tidal shoal often causes local longshore transport reversals, resulting in onshore or alongshore transport of sediment. On the downdrift side of the inlet, direction refracted waves produce a longshore current flowing toward the inlet, that is contrary to the prevailing longshore drift.

The morphology of Ocean City Inlet's ebb shoal is crescentic in shape, and offset to the south. The shoal extends approximately 1.3 km offshore, and is approximately 1.9 km wide. Table 5 shows grain size analysis from nine vibracores extracted from the ebb shoal in 1986 (Anders and Hansen 1990). Sand thicknesses of 6 m exist on the eastern end, with 3 m portions on the north side of the ebb shoal. The mean grain size for the entire shoal was 2.1 phi. In general, the subsurface sediment structure was one of sands overlaying clays or peat, sand sizes ranging from 0.59 phi to 3.38 phi, and moderately to poorly sorted material. These shoal mean grain sizes are similar to mean grain sizes found on the adjacent beach of Ocean City (1.84 phi, standard deviation of 1.22 phi).

An attempt by the authors to estimate the equilibrium storage volume of Ocean City Inlets ebb shoal, as it relates to inlet throat cross sectional area (A) and tidal prism (P), was calculated using equations ($V = a'A^b$) and ($V = a'P^b$) from Walton and Adams (1976), where:

V = ebb shoal volume

a, b = correlation coefficients derived through linear regression of data from 44 U.S. inlets (Walton and Adams 1976).

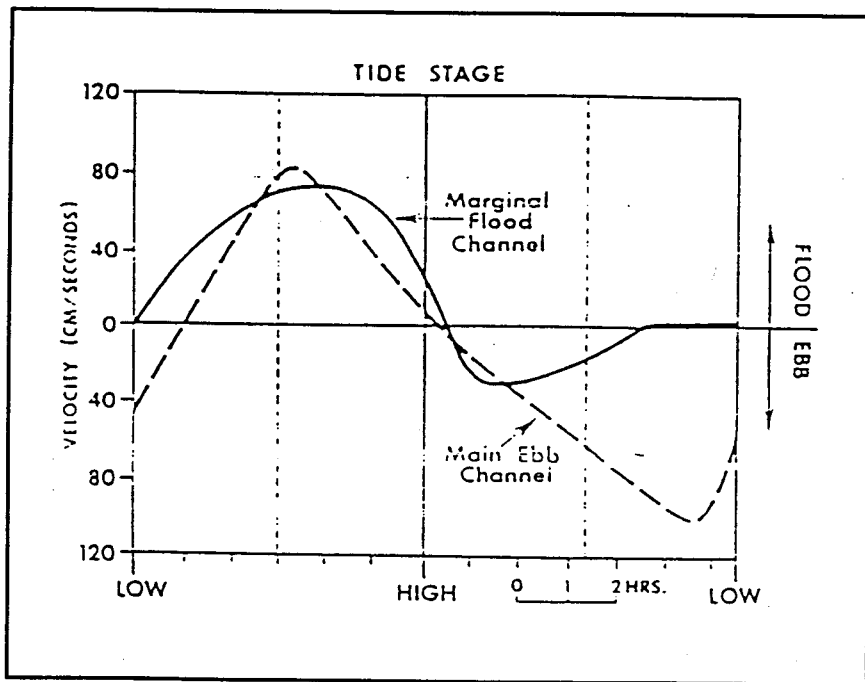


Figure 35. Time velocity asymmetry for typical inlet, note maximum tidal stage velocities do not occur at mid-stage (After Postma 1967, From Hayes 1980).

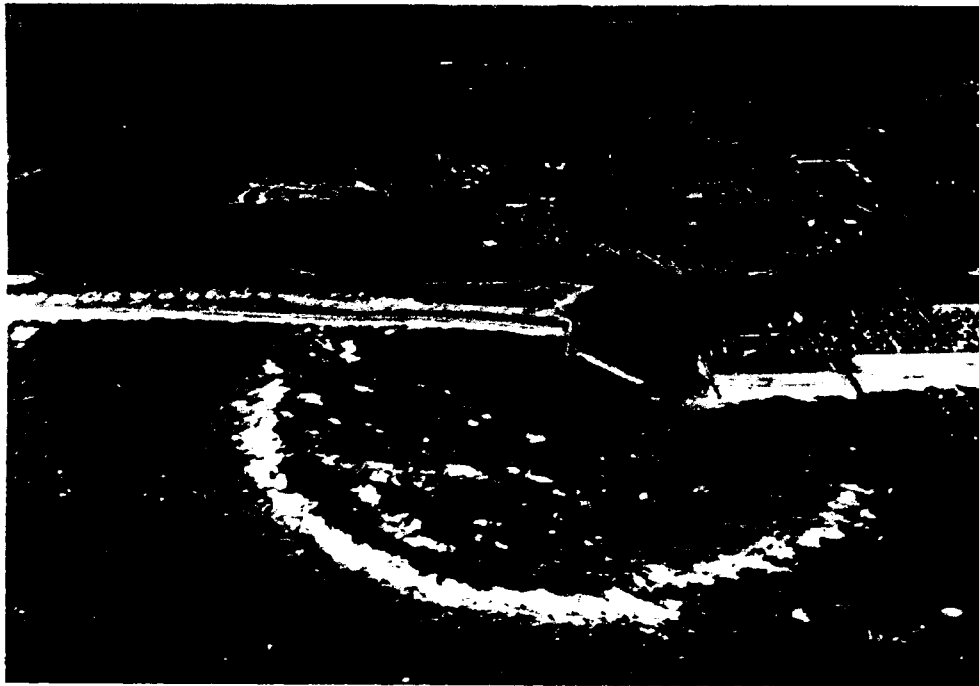


Figure 36. Photograph of ebb shoal in October 1991. Note shielding benefit as waves break along outer edge of shoal.

Table 5

Summary of Vibracore Analysis for Ebb Shoal (from Anders and Hansen 1990)

Core No.	Location State Plane (ft)	Descriptive Location and Depth (m)	Max. Sand thickness (m)	Mean Size (phi)	Stand. Deviation (phi)	Skew (phi)	Mud %
1	180,200 1,346,900	S. Flank 6.7	2.2	2.84 Fine	0.66	-0.6	4
2	182,500 1,350,700	N. Flank 4	4	2.31 Fine	0.56	-1.0	0.2
3	182,350 1,347,800	N. Flank 4	1.5	1.80 Medium	0.80	-0.6	0.1-1.9
3b	"	"	1.8	2.24 Fine	1.05	0.0	3.2-17.6
4	180,050 1,350,050	S. Flank 7	3.4	2.11 Fine	0.70	-0.2	0.1-0.7
5	181,500 1,349,200	Crest 2.1	2.7	1.63 Medium	0.73	-2.0	0.0
6	182,500 1,349,00	N. Flank 5.2	3.9	1.42 Medium	0.70	-0.2	0.0
7	180,200 1,352,00	S. Trough 9.1	1.4	2.65 Fine	0.78	-1.2	2.3
"	"	"	1.6	2.35 Fine	1.03	-1.0	10.6-11.4
8	183,550 1,350,800	N. Trough 6.7	5.3	1.89 Medium	0.59	-0.6	0.0-0.1
9	182,600 1,350,300	N. Flank 4.6	4.6	1.49 Medium	0.69	-0.2	0.1

CROSS SECTIONAL AREA METHOD ($V = a'A^b$)

V = ebb shoal volume

A = Cross Sectional Area = 13,250 ft² (Dean et al. 1979)

V = 40.7A^{1.28}

A = Cross Sectional Area = 13250 ft² (Dean et al. 1979)

P = Tidal Prism = .81 x 10⁹ ft³ (Dean et al. 1979)

Coast Type = Moderately Exposed

Ebb Shoal Volume = 5.8 x 10⁶ m³

TIDAL PRISM METHOD ($V = a'P^b$)

V = ebb shoal volume

P = Tidal Prism = .81 x 10⁹ ft³ (Dean et al. 1979)

V = 10.5 x 10⁻³P^{1.23}

Ebb Shoal Volume = 7.2 x 10⁶ m³

The volume of the ebb shoal reached approximately 6.0 x 10⁶ m³ in 1978. Averaging the above ebb shoal volumes equates to approximately 6.5 x 10⁶ m³. Using Walton and Adams (1976) criteria, these two estimates reveal that the shoal has reached, or nearly achieved, its "*equilibrium capacity*" to store sediment as early as 1978. Based on this information, one would expect sediment to bypass the inlet system and move downdrift to Assateague Island, assuming there are no changes to tidal prism or inlet cross sectional area.

Inlet Stability

Newly formed tidal inlets may remain permanently open. Others are ephemeral in nature, originating during storms and eventually closing due to a combination of natural coastal processes and sediment supply. The term "stable inlet" should be understood in a relative sense, because no absolutely stable inlet exists. Escoffier (1977), states that "the stability of a tidal inlet should be interpreted as a "dynamic stability" by which acting forces maintain a situation characterized by relatively minor changes in inlet geometry including location, planform, cross sectional area and shape. It is a matter of importance to be able to determine the hydraulic characteristics which control the stability of tidal inlets."

Tidal inlet hydraulics is a combination of time-dependent factors that include (1) ocean processes (wind, waves, storm surge, tide), (2) nearshore conditions (longshore currents, man-induced modifications to the shore), (3) geometric characteristics of the bay and inlet channel, and (4) inlet sediment characteristics (Mayor-Mora 1977). The movement of water through an inlet is caused primarily by tides. As the ocean rises and falls with the tide, the flow of water through

an inlet causes the water level in the bay or lagoon to also rise and fall. These differences between the ocean and bay water levels determine the amount of flow through the inlet. Over a tidal cycle, tidal flows start up, shut down and reverse, creating a set of diagnostic flow patterns. Dean and Walton (1975) describe an inlet's shape as "an idealized water nozzle". "Ebb flow patterns expel strong, steady unidirectional turbulent jets of water, out and over the ebb shoal, while the flood flow patterns are more streamlined. On the ebb flow, (at low water levels) the momentum of the existing water forms a jet directed seaward. The ebb jet's higher velocities carry sediment considerable distances seaward. Eddies generated by this jet exiting seaward move water toward the inlet on ebb tide, through the marginal flood channels. On flood tide, water converges toward the inlet on all sides" (Figure 37).

In determining the stability characteristics of inlets, much effort has been put into establishing empirical relationships. For example, inlet stability was defined by the relationship between inlet cross section (A_c) and tidal prism (Ω) in the bay (O'Brien 1931, 1969). Jarrett (1976) used inlet cross-sectional area/tidal prism relationships for inlets on the Atlantic, Gulf of Mexico, and Pacific coasts. He concluded that the different coasts exhibited different stability characteristics as a result of differences in the tidal and wave characteristics. From O'Brien and Jarrett's work, an equilibrium condition for inlet size (cross sectional area) is determined primarily by magnitude of the tidal prism, assuming that adjustments to inlet cross-sectional area are a response to the balance between bed shear stress in the inlet and changes in the magnitude of littoral drift. Therefore, an increase in tidal prism tends to enlarge the inlet cross section by increased tidal scour. Conversely, a decrease in tidal prism (i.e. reduced tidal flows) reduces inlet cross section by permitting accumulation of littoral sand in the inlet. Bruun and Gerritsen (1959), studied ratios between a natural inlet's tidal prism (Ω) and annual gross littoral drift (M_{total}) to provide some measure of an inlet's ability to naturally remain open. Dividing tidal prism by gross littoral drift, resulted in values greater than 200 indicating good stability characteristics, values between 100-200 gave fair stability, and values below 100 were classified as an inlet with poor stability. Inlets with high ratios are providing sufficient tidal flow velocity (from both ebb and flood tidal cycles) to prevent substantial sediment buildup within the inlet. An inlet with a lower ratio has a greater possibility of *sanding itself in* due to the inability of tidal flow to overcome and remove accumulated sediments within the inlet throat section. The influence of jetties on an inlet, tend to increase the stability characteristics by reducing littoral drift quantities into the inlet.

Escoffier (1940,1977) suggested the concept of maximum current velocity controlling inlet cross sectional size. He illustrated that there is a maximum velocity curve (V_{max}) at some critical cross section area of the inlet. This information implies that inlet areas greater than the critical area will result in a decrease in current velocity, causing deposition and area reduction, until stabilization occurs at the equilibrium area. Conversely, inlet areas less than the critical area will result in increased velocities and scour until equilibrium is attained at the appropriate cross sectional area. A hydraulically unstable inlet will either continuously scour its throat until the critical flow area is achieved, or it will continuously shoal until its critical flow area is achieved or the inlet closes.

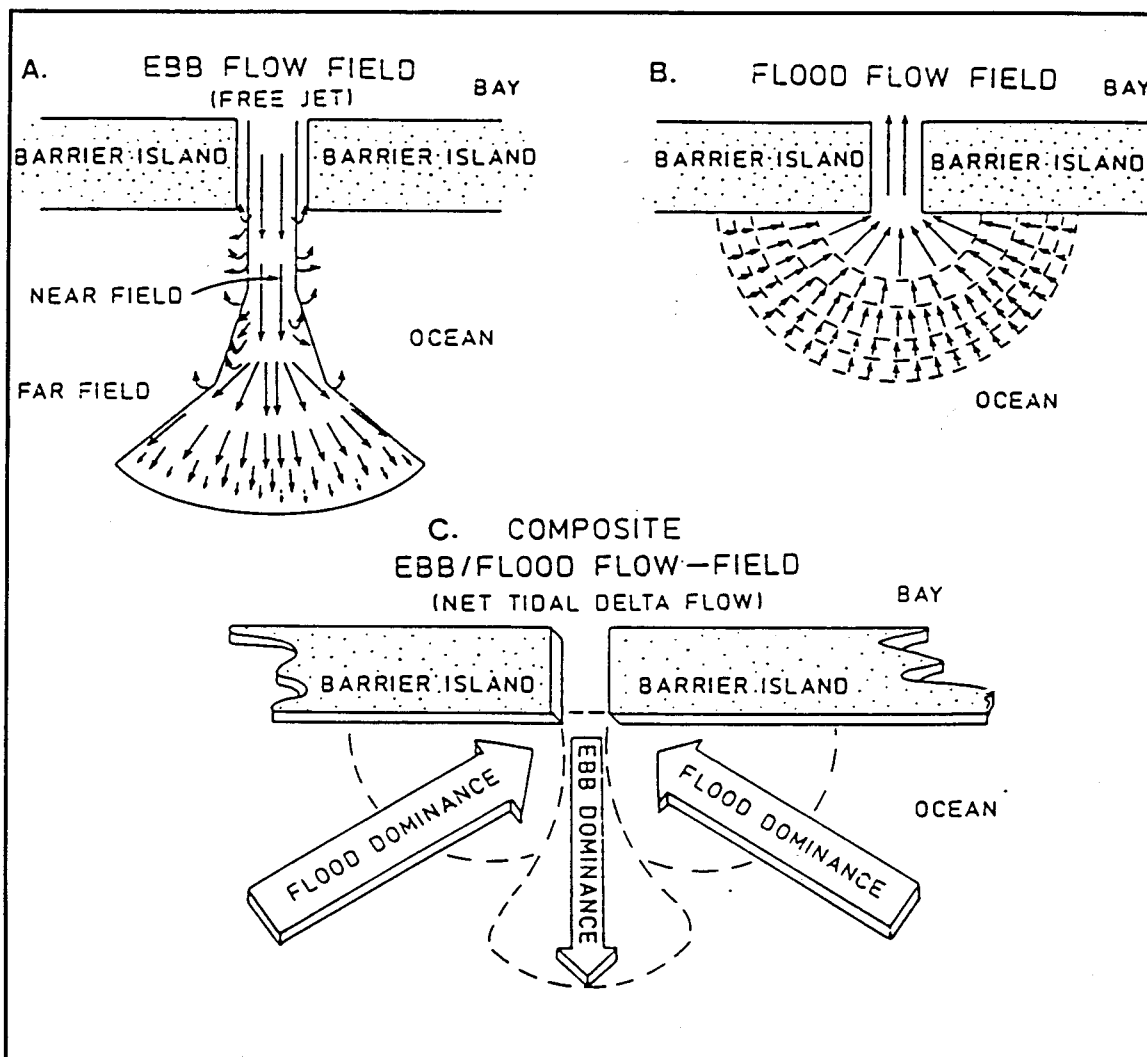


Figure 37. Idealized ebb and flood flows over a typical tidal cycle (From Oertel 1988).

In order for Ocean City Inlet to remain permanently open (i.e. increased stability), jetties had to be constructed along the north and south sides of the inlet, blocking longshore littoral drift from entering the inlet. These jetties prevented longshore littoral drift from choking the inlet channel with sediment. The jetties served to entrain the tidal flows, thereby increasing the flushing capability of the inlet.

Dividing Ocean City Inlets tidal prism by gross longshore littoral drift (Ω/M_{total}) results in a stability value of 114, where:

Ocean City inlet's tidal prism = $2.43 \times 10^7 \text{ m}^3$

This value is 10 percent and 23 percent larger than the values obtained using O'Brien's (1969) and Jarrett's (1976) formula, respectively (Dean et al 1979).

Gross longshore littoral drift = to Net littoral drift, because Net is a large percent (25 percent) of gross littoral drift = $2.128 \times 10^5 \text{ m}^3$ (Douglas 1985)

This value would indicate an inlet with fair *natural* stability characteristics. If Ocean City Inlet hadn't been artificially stabilized, this classification would be reasonable based on historic records of numerous inlets opening and closing along this stretch of shoreline. For example, Sinepuxent Inlet remained open approximately 8 years before closure. However, due to entrainment by the two jetties, the inlet appears to have considerable flushing capacity (increasing stability as compared to being left naturally open) as evidenced by the enormous ebb shoal, and the fact that the ebb shoal platform has almost always maintained a navigable depth. Only minor dredging events in the main inlet throat section have been needed. Examination of dredging locations for the inlet reveal that most dredging activities have taken place along the western junction of Sinepuxent and Isle of Wight Bays (i.e. Commercial Fish Harbor) (Wicker 1974).

A comprehensive study of shoaling within Ocean City Inlet was conducted by Wicker (1974) and Dean et al (1979). The results from these reports will be used extensively in this section to help explain water movement and circulation, tidal elevations and velocities, and sediment transport patterns within Ocean City Inlet. Dean et al. (1979), used a numerical model to calculate the tidal prism and distribution of flow within the back bay areas. Model results disclosed that predominant flow was towards the north (85 percent) into Isle of Wight Bay, with only 15 percent moving southward into Sinepuxent Bay. Symmetrical, or slightly asymmetrical flood patterns are assumed due to the outer jetty configuration. The north jetty extends slightly farther seaward than the south jetty. Due to the northern predominance of flow from Isle of Wight Bay, scouring along the inner portion of the south jetty has occurred in the past. Evidence of scouring was first recognized as early as 1937-1938. The scour hole had increased from 2.4 - 3 m during 1930-1940, and to 12 - 15 m deep in 1982 (USACE 1983). The inlet channel does shift from time to time between the two jetties. Evidence of this shifting is found on National Oceanic and Atmospheric Administration (NOAA) chart no. 1220. This chart includes the note "Ocean City Inlet - The channel is subject to continual change. The buoys are not charted because they are frequently shifted in position". Presently, Ocean City's navigational channel is directed north, northeast of the outer jetty boundaries.

Long term comprehensive current velocity data through the inlet is scarce. Some current velocity measurements were acquired by Wicker (1974), Dean et al (1979), and USACE (1983). Results from these studies show that current velocities within the main inlet channel, for both ebb and flood tide, averaged approximately 1.06 m/sec. Spring tidal flows as a result of storm occurrences would be expected to increase these velocities appreciably. Bruun (1978) states "maximum current velocities during a spring tidal cycle are generally about 10-15 percent greater than during a mean tidal cycle". A 15 percent increase from 1.1 m/sec during a spring tide, yields approximately 1.2 m/sec velocity. Therefore, spring tidal velocities associated with a significant storm event, may result in current velocities of 1.2 m/sec. According to USACE (1991), current velocities within the range of .91-1.2 m/sec are capable of suspending bed and bank material with grain sizes of fine to medium sand. Sediments within this range are found within the inlet. Presently, the ebb and flood tidal flow seems to be in a relative state of equilibrium with sand supply (i.e. flow is sufficient to prevent shoaling in this channel section). Throughout the history of the main inlet channel proper, there has never been less than 3 m depth over a distance of 61 m, the active project dimensions (Wicker 1974).

Engineering Impacts

Newly created tidal inlets, through artificial or natural means, disrupt the continuity of local sediment transport processes along the coast. Constant alongshore currents move sand towards the inlet at all times. Sediment accumulation may occur both landward (flood tidal delta) and seaward (ebb tidal delta) of the tidal inlet at the expense of sediment in the littoral transport system. In some cases, artificial stabilization (i.e. jetties) of these tidal inlets is necessary for these natural or artificial shoreline breaches to remain open. Unless stabilized, these inlets tend to be ephemeral in nature, opening and closing over time in response to changing nearshore coastal conditions. Once stabilized, profound changes to the adjacent shorelines may occur (both updrift and downdrift) through interruption and subsequent impoundment of littoral drift material (Leatherman 1979, Dean 1988, and Oertel 1988). These adverse effects to adjacent shorelines are primarily caused by interruption of the natural bypassing mechanisms or "sand sharing system" of an inlet. Dean (1988), describes this system as "the interconnectiveness and equilibrium of the sand bodies comprising the inlet, ebb tidal shoal, and adjacent shorelines".

The three engineering activities that have had the greatest impact on northern Assateague Island (as pertaining to longshore sediment transport) have been; (1) inlet stabilization through construction of jetties, (2) sand tightening of the south jetty, including construction of headland breakwaters and (3) beach nourishment of Ocean City's shoreline. The first and most important of these activities was interruption of longshore sediment transport caused by inlet formation (1933) and the subsequent stabilization with jetties (1935), resulting in dramatic shoreline changes on either side of the inlet. All or part of the sediment reaching the jetty/inlet system by longshore drift was either trapped along the northern jetty or moved to the inlet associated shoals, both landward and seaward.

The second engineering impact was rehabilitation of the south jetty with pre-cast interlocking blocks and raising the crest elevation from a +1.4 m to +2.3 m NGVD. This structural improvement prevented sediment from entering the inlet by wave overtopping and structure permeability. Wave refraction around the ebb tidal delta reversed littoral drift movement along the shore, moving sediment north towards the inlet throat. This structural rehabilitation helped trap sediment along the northern portion of Assateague Island. Before this construction, sediment migrated over and through the south jetty settling within the inlet and along the northwestern portion of the island. However, once sediment was prevented from moving over and through the south jetty, erosion of the northwestern portion of Assateague Island was predicted. Three headland breakwaters were constructed along the northwestern shoreline to stabilize this area from scouring effects of tidal currents and wave action.

Lastly, beach nourishment activities since 1988 may be playing an important role in slowing erosion rates adjacent to northern Assateague. Intuitively, one would suspect that any additional sediment input directly to this area would provide additional material to the adjacent beaches. Net littoral drift from north to south would eventually carry sediment to and/or around the inlet (natural sediment bypassing) to nourish Assateague Island. The exact impact of beach nourishment activities on northern Assateague is beyond the scope of this paper. However, the authors feel there is a strong possibility that the sediment placed on Ocean City beaches is having a positive impact on the erosion rates of northern Assateague Island.

Bypassing at Natural and Jettied Inlets

One mechanism of normalizing littoral drift movement downdrift of "littoral barriers" (i.e. inlets, jetties) is by natural sediment bypassing. For example, downdrift of an inlet, the littoral drift system may be undernourished resulting in high erosion rates along the barrier island. The inlet's ebb and flood tidal delta acts as a sink for sediment passing around the jetty or, due to strong ebb tidal currents, forces sediment offshore and out of the active littoral drift zone. Eventually, natural sediment bypassing may occur, restoring much needed sediment to downdrift beaches. Natural sediment bypassing mechanisms of tidal inlets have been well documented (Bruun and Gerritsen 1959, Humphries 1977, FitzGerald 1988, Dean 1988, and Tomlinson 1991). The two main principles of tidal inlet bypassing are offshore bar bypassing and tidal flow action (Figure 38). These processes, more often than not, occur in conjunction with each other. Offshore bar bypassing occurs when the seaward shoal acts as a "subaqueous bridge" upon which sediment is transferred across the inlet to the downdrift shoreline. Tidal flow bypassing functions by the action of ebb and flood currents moving sediments into and out of the tidal inlet. These sediments may eventually be deposited in the offshore bar, moved alongshore across the bar by wave action, and returned to the shore downdrift of the inlet. Whether predominant bar bypassing mechanisms transcend tidal flow bypassing depends on the ratio (r) between littoral drift and tidal flow (Bruun and Gerritsen 1959). If the predominant littoral drift (M_{mean}) is expressed in cubic units per year, and maximum tidal flow under spring conditions, in the same cubic units per second

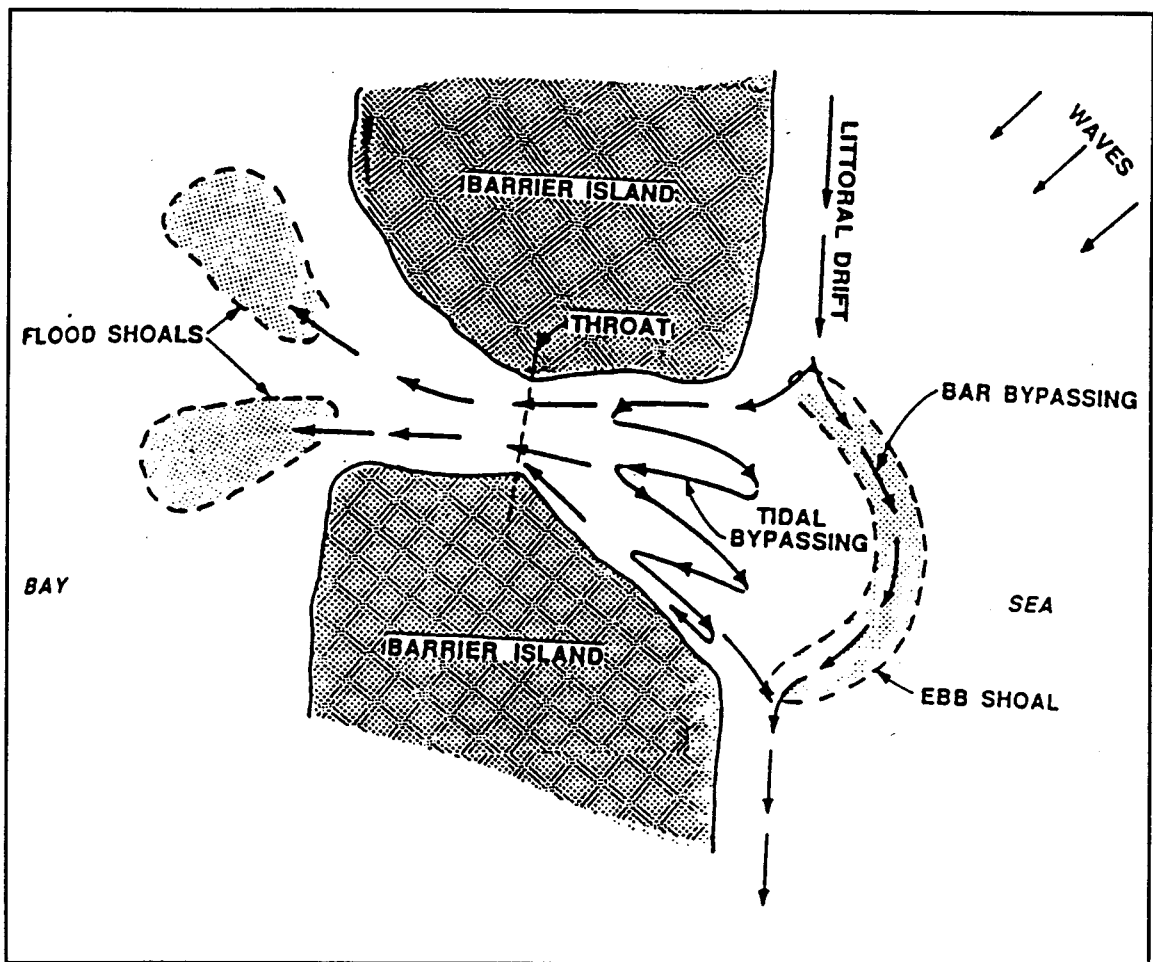


Figure 38. Two main principles of tidal inlet bypassing - offshore bar and tidal flow (From Mehta et al 1989).

(Q_{max}), then bypassing is described by the ratio: $M_{mean}/Q_{max} = r$. Large ratios indicate predominant bar bypassing ($r > 200-300$). Small ratios indicate predominance of tidal flow bypassing ($r < 10-20$). Sediment transport processes may occur more rapidly during storm periods when wave conditions, tidal elevations, and currents are increased (Hite 1924, FitzGerald 1988, Webb et al. 1989, Douglas 1985). Figure 39 depicts tidal inlet processes operating during a flood cycle of a storm.

Nearshore processes that occur adjacent to natural inlet entrances provide the basis for understanding sediment transport processes around modified inlet entrances. The manner in which inlets bypass sand via their ebb-tidal shoals controls the rate and location of sediment nourishment to the downdrift shoreline (FitzGerald 1982). FitzGerald (1988) gives three examples of sediment bypassing mechanisms that occur along mixed energy, natural inlet entrances; inlet migration and spit breaching, ebb-tidal delta breaching, and stable inlet processes (Figure 40). Inlet migration and spit breaching occur due to a downdrift migration of the channel thalweg by updrift accumulation of sand and wave generated currents, eventually rendering the inlet hydraulically inefficient. The spit breaches creating a more direct flow offshore. The abandoned ebb channel becomes filled and migrates landward welding itself to the downdrift shoreline. Ebb-tidal delta breaching (model 3, FitzGerald 1988) is somewhat similar in its bypassing mechanisms as in FitzGerald's model 1, except that the inlet is characterized as having a stable inlet throat position. Over time, migration of the main ebb channel occurs in a downdrift direction through accumulation of sediment updrift of the inlet (i.e. dominant direction of longshore sediment transport). Again, the inlet becomes hydraulically inefficient, causing a break seaward through an existing spillover lobe channel. Bypassing occurs through swash bar migration and wave induced sediment transport across the swash platform. The stable inlet (model 2) involves a stable inlet throat position and a main ebb channel that does not migrate across its ebb tidal delta. FitzGerald (1988), describes sand bypassing by stable inlets through "formation, landward migration and welding of large bar complexes to the downdrift shoreline". These bar complexes are the result of the stacking and coalescing of wave built swash bars situated across the ebb tidal delta platform. These swash bars move onshore as a result of net landward transport over the swash platform. Figure 41, illustrates the mechanisms involved in net landward transport of sediment towards the adjacent shoreline. Breaking waves along the terminal lobe will currents and enhance flood currents. As these swash bars move landward, their rate of onshore migration decreases due to movement up the beach slope, and greater intertidal exposure (Figure 42). The location of this bar attachment on the downdrift shoreline is directly affected by (1) inlet size, (2) wave versus tide dominance, and (3) channel orientation (FitzGerald 1982). These three factors are illustrated in Figure 43. Large inlets naturally have larger bar/shoal complexes than smaller tidal inlets. Inlets found along coasts with high longshore transport rates should have a greater capacity to bypass sand due to the greater volume of sand available for bypassing. Wave and tide dominated inlet systems transfer material to the downdrift beach by their respective dominant process. Wave dominated shorelines move sediment around the ebb shoal (terminal lobe) and across the ebb shoal platform. Tide dominated inlets bypass sediment through tidal currents and periodic bar welding events. An inlet's main ebb channel orientation controls the distance from the inlet at which bar complexes weld themselves to the shoreline. A straight seaward ebb channel orientation allows for bar complexes to attach

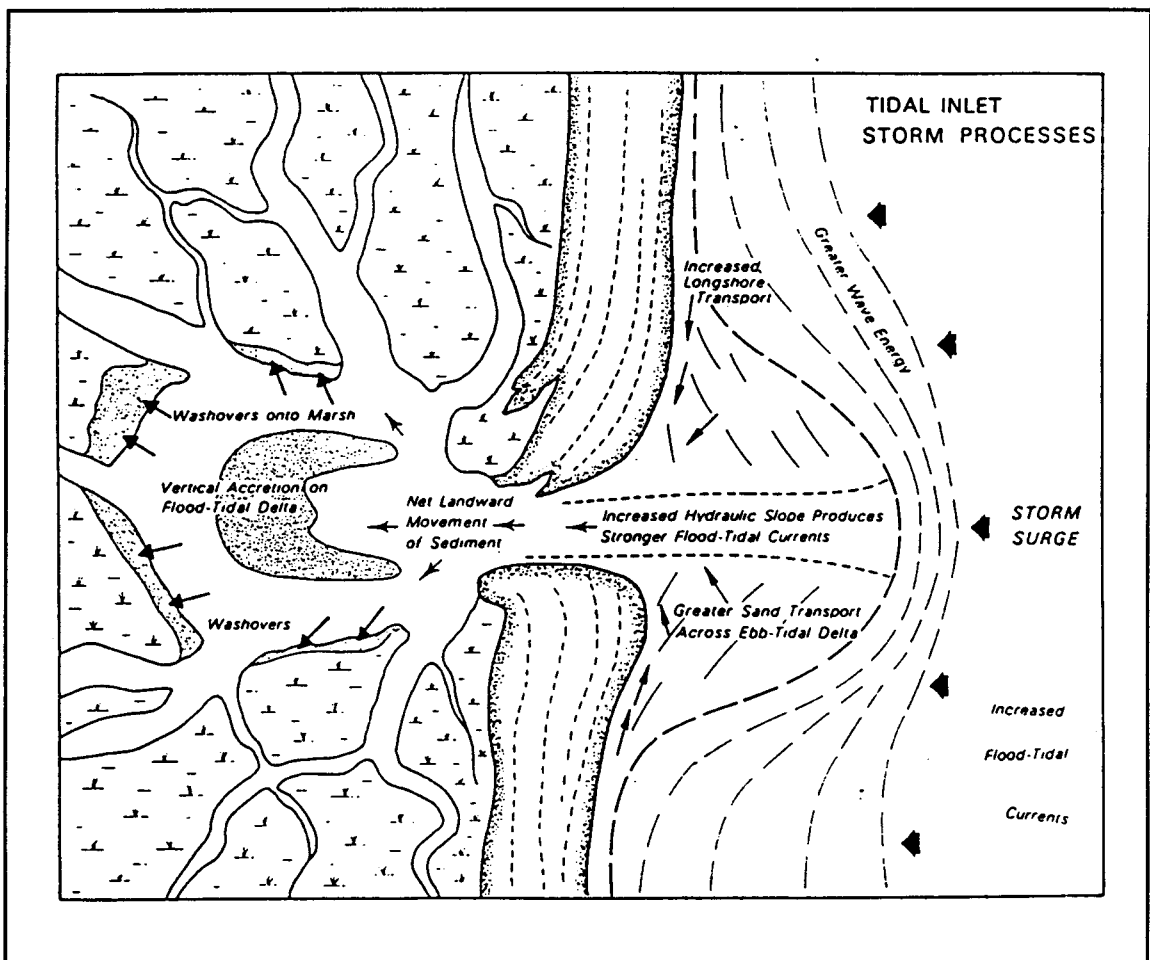


Figure 39. Storm induced inlet processes on a flood tide (From FitzGerald 1988).

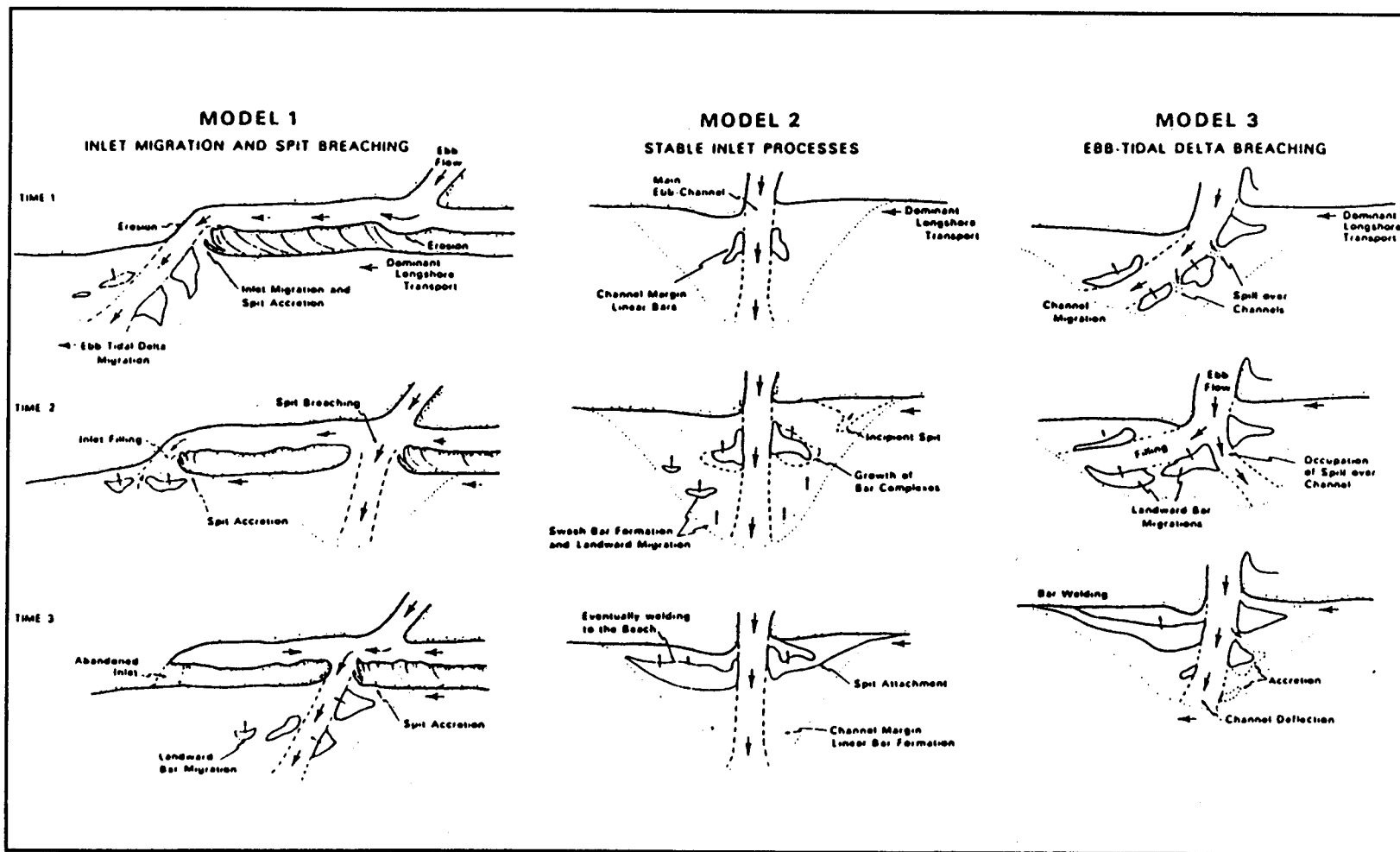


Figure 40. Three sediment bypassing mechanisms along mixed energy, natural inlet entrances (From FitzGerald 1988).

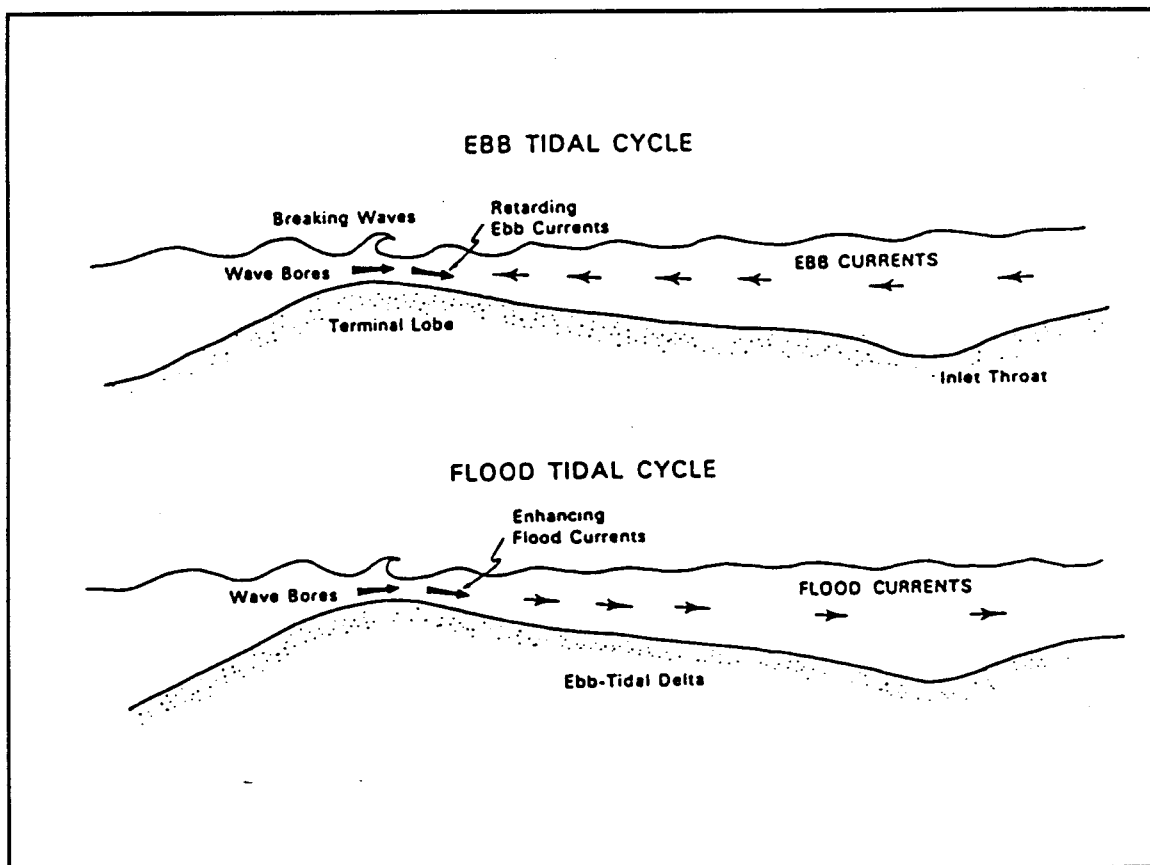


Figure 41. Mechanisms involved in net landward transport over the ebb shoal platform (From FitzGerald 1988).

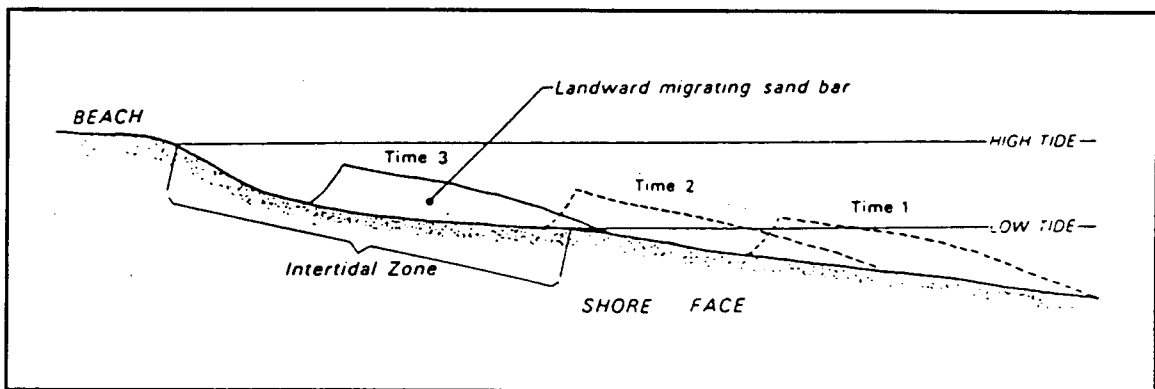


Figure 42. Onshore movement of swash bars from wave action (From FitzGerald 1988).

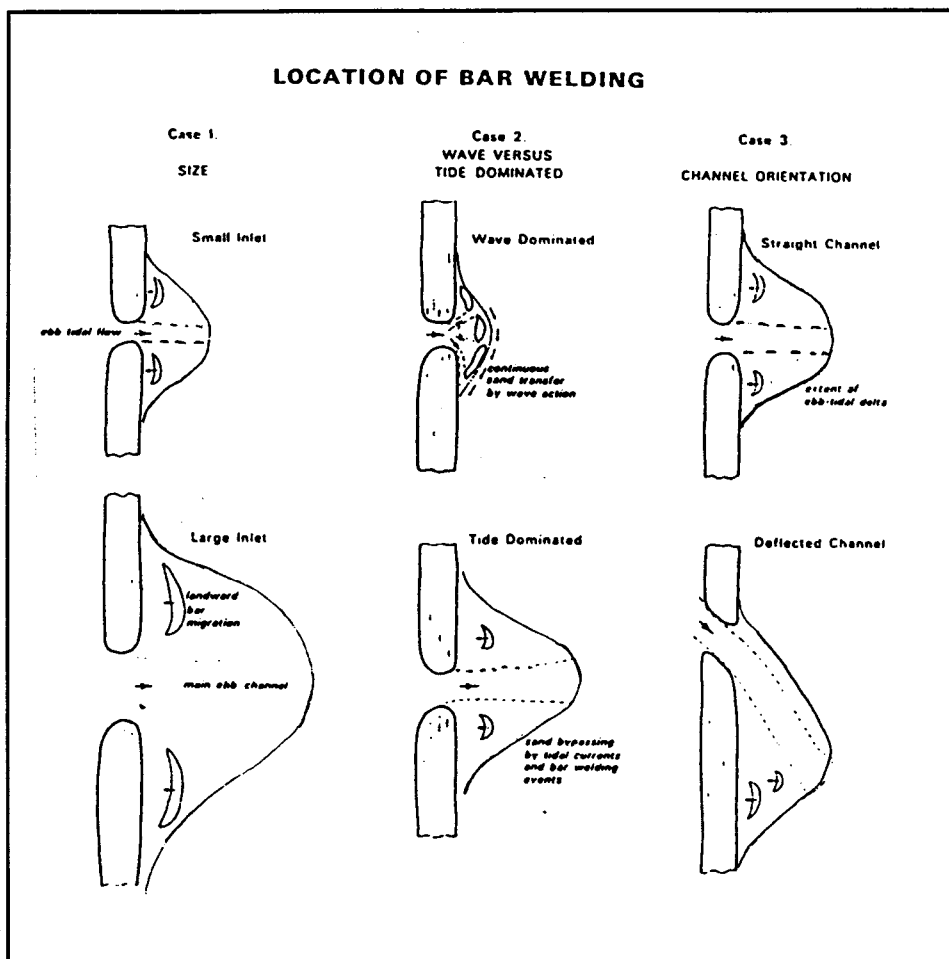


Figure 43. Three factors that affect the location of bar attachment (From FitzGerald 1982).

relatively close to the inlet. Downdrift deflections of the main ebb channel result in bar complexes welding onshore some distance away from the inlet. The attachment point of these bar complexes along the downdrift barrier shoreline creates a distinctive bulge in the shoreline where these bar complexes attach themselves to the beach (FitzGerald 1982).

In the past, sediment bypassing downdrift of stabilized inlets has been predominately through artificial means (Dean 1988, Stauble et al 1987, Clausner et al 1991). Traditionally, this method required physically and/or hydraulically transferring material from the depositional side of the inlet to the erosional side by conventional dredges or fixed plant operations. These bypassing methods become necessary due to the destabilizing effect that jetties and other structures have on local sediment transport processes. An example of a permanent bypassing system is found north of Ocean City at Indian River Inlet, Delaware. This bypassing plant has demonstrated its capability to transfer approximately 76,000 m³ of sediment from the accretional southern side to the erosional northern side of the inlet (Clausner et al 1991).

When a newly created or existing unstable inlet is "improved" for navigational purposes by the addition of jetties, these jetties form a barrier to longshore transport of littoral drift. Typically, where longshore sediment transport predominates in one direction, jetties will cause deposition on the updrift shoreline and erosion on the downdrift shoreline. Jetties may also have the capability of displacing the ebb shoal farther seaward in deeper water. This usually results in a considerable ebb shoal volume increase, and disruption of equilibrium conditions which may have existed prior to stabilization and/or extension by jetties (Bruun and Gerristen 1959, Dean 1988, Tomlinson 1991). When an inlet is stabilized by jetties, the associated offshore and inshore bathymetry change until an equilibrium is approached with the predominant currents and incident wave climate (Tomlinson 1991). Typically, jetties are constructed of quarry stone, and flank both sides of the desired channel alignment, and extend from the shoreline to a depth usually governed by the desired depth of the entrance channel (Seaburgh 1983). Jetties provide the following functions: (1) block entry of littoral drift into the navigation channel, (2) stabilize a migrating inlet, (3) serve as training walls for inlet tidal currents, and (4) increase velocity of tidal currents, flushing sediments from the channel (USACE 1984).

The presence of a large ebb shoal and recent reductions in erosional rates for northern Assateague Island indicate that conditions are favorable for offshore bar bypassing at Ocean City Inlet. Utilizing Bruun and Gerritsen (1959) ratio of $M_{\text{mean}}/Q_{\text{max}}$ where:

$$M_{\text{mean}} = \text{net longshore littoral drift} = 212,800 \text{ m}^3$$

$$Q_{\text{max}} = \text{maximum tidal flow under spring conditions} = 1,431 \text{ m}^3$$

yields a "r" value of 148, suggesting offshore sediment bar bypassing as the predominant transfer mechanism ($r > 200$ –300, with bar formation) as compared to tidal flow bypassing ($r < 10$ –20, little or no bar formation).

5 Methods

Shoreline Position Change

A computer-based shoreline mapping methodology (McBride et al. 1991, Byrnes et al. 1991, 1993), within a GIS framework, was used to compile and analyze changes in historical shoreline position between 1929/33 and 1989 for Fenwick and Assateague Islands near Ocean City Inlet, Maryland. The purpose of this effort is to quantify changes in shoreline position using the most accurate data sources and compilation procedures to characterize the morphological evolution of the Ocean City tidal inlet complex. Because this information addresses a critical component of a broader study of morphologic evolution of Ocean City Inlet ebb-tidal delta and the impacts of jetty construction on adjacent shorelines, emphasis was placed on data accuracy and potential error estimates for gauging the significance of results.

Data Sources

Two data sources were utilized for assessing spatial and temporal changes in shoreline position. These are National Ocean Service (NOS) topographic maps and near-vertical aerial photography. NOS Topographic maps (T-sheets) are produced by the US Coast and Geodetic Survey (USC&GS) (now National Oceanic and Atmospheric Administration - NOS). Because this agency is responsible for surveying and mapping topographic information along the coast, T-sheets have been used in the study of coastal erosion and protection, and frequently in courts in the investigation of land ownership (Shalowitz 1964). Most of these maps are planimetric in that only horizontal position of selected features is recorded; the primary mapped feature is the high-water shoreline. Scales of T-sheets are generally 1:10,000 or 1:20,000, although others exist. These larger-scale products provide the most accurate representation of shoreline position other than direct field measurements using engineering or satellite surveying methods. For this study, NOS T-sheets with survey dates of 1849, 1850, 1929, 1933, 1942, 1961, 1962, and 1976 were digitized. A 1908 shoreline, originally digitized from a NOS T-sheet by USACE, was digitized from USACE compilation maps because the T-sheet was not available. A March 1980 shoreline, photo-interpreted by NOS, was also digitized from these compilation maps.

Since the 1920's, aerial photography has been used to record shoreline characteristics in many coastal regions. However, these data cannot be used directly to produce a map. Aircraft tilt and relief may cause serious distortions that must be removed by rectification. A number of graphical methods and computational routines exist for removing distortions inherent in photography (Leatherman 1983; Anders and Byrnes 1991). Orthophotoquads and orthophotomosaics are photomaps made by applying differential rectification techniques (stereoplotters) to remove photographic distortions. Ease of data collection and the synoptic nature of this data source

provide a significant advantage over most standard surveying techniques. Aerial photography taken April 12, 1989 was interpreted and digitized for this task.

Digitizing Shoreline Position Data

A detailed methodology involving computer cartography, CADD, and GIS software has been developed for digitizing shoreline position information from maps and aerial photographs (Byrnes et al. 1991, Byrnes and Hiland, 1993). This methodology was used for this task and is briefly described below.

When digitizing cartographic information, care must be taken to exactly represent the map being digitized. This means using computer cartography software to set up a file with a projection, datum, and coordinate system identical to the map. Digitizer setup (tying digitizer and file coordinates to real world coordinates represented on the map) is performed using the most accurate points available on the map and using well-spaced points to improve the geometry of the transformation matrix. Shoreline position is carefully digitized using a high precision (± 0.1 mm) digitizer and cursor. After all maps are digitized, checked for errors, and corrected, they are converted to a common horizontal datum, projection, and coordinate system (Universal Transverse Mercator Zone 18, NAD 83). This is critical for accurately overlaying and comparing shorelines of different dates.

Aerial photography is rectified to a stable base map (usually NOS T-sheets) using a Zoom Transfer Scope to optically remove distortions. Shoreline position is traced onto a stable film, along with longitude-latitude points from the map. This photomap is then digitized using the same techniques as for cartographic data. The files for digitizing photomaps are set up to match the cartographic parameters of the maps used for aerial photograph rectification.

Potential Error Estimates

It is important that all available procedures be used to capture map data as carefully as possible; however, no matter how cautious the approach, a certain measure of error will be retained in all measurements of digitized horizontal position. Potential errors are in terms of accuracy and precision, and both types of error should be evaluated to gauge the significance of calculated changes relative to inherent inaccuracies.

Shoreline measurements obtained from historical maps can only be as reliable as the original maps themselves. Accuracy depends on the standards to which each original map was made, and on changes which may have occurred to a map since its initial publication. Field and aerial surveys provided the source data used to produce shoreline maps. For T- and H-sheets at a 1:10,000 scale, national standards allow up to 8.5 m of error for a stable point (up to 10.2 m of error at 1:20,000), but the location of these points can be more accurate (Shalowitz 1964, Anders

and Byrnes 1991, Crowell et al. 1991). Non-stable points are located with less accuracy; however, features critical to safe marine navigation are mapped to accuracy stricter than national standards (Ellis 1978). The shoreline is mapped to within 0.5 mm (at map scale) of true position, which at 1:10,000 scale is 5.0 m on the ground.

Potential error considerations related to field survey equipment and accurate mapping of high-water shoreline position include accuracy of measured distances (± 1 meter), determining position of the survey equipment (± 2 to 3 meters), identification of the high water line (± 3 to 4 meters), and sketching of the shoreline between measured points (up to ± 5 meters) (Shalowitz 1964).

In addition to survey limitations listed by Shalowitz (1964), line thickness and cartographic errors (relative location of control points on a map) can be evaluated to provide an estimate of potential inaccuracy for source information. Line thickness, due to original production and photo-reproduction, was no greater than 0.3 mm, or 6 m ground distance at 1:20,000 scale.

Beginning in the late 1920's, NOS T-sheets were compiled from rectified aerial photography. In addition to inherent cartographic error sources, a potential error of ± 5 meters is incurred because of the remote interpretation of shoreline position. Another source of error relates to equipment and operator accuracy and precision. The absolute accuracy (accuracy and precision) of the digitizing tables used for this study is 0.1 mm. At a scale of 1:10,000, this converts to ± 1 m. Furthermore, the precision with which an operator can visualize and move the cursor along a line can lead to much greater errors (Tanner 1978). Fortunately, improper tracking associated with shoreline digitizing generally is random and may be dampened when averaged over finite distances of shoreline. To evaluate the magnitude of operator error associated with digitizing shoreline position, at least three repetitive measurements should be compared. For this study, the average error incurred using this procedure for a 1:20,000 scale map was about ± 2 m.

Quantifying Shoreline Change

Once shoreline position data are compiled accurately, spatial and temporal changes can be quantified using manual or automated procedures. The manual technique involves making measurements at regular intervals from overlay maps or from the computer screen and tabulating this information for assessing trends. Not only is this approach time consuming, but, if analog map overlays are used, another level of inherent error is included in the measurements because the composite map is a second-generation product at scale. Most current procedures applied for quantifying changes in shoreline position involve some method of digital data comparison.

For this study, the Automated Shoreline Analysis Program (ASAP) was used to quantify shoreline change at a 50-m longshore interval. The first step of the procedure is to identify segments of the coast with similar shoreline orientations. Next, digital data stored in an Intergraph MicroStation design file are imported by a software routine that prepares the information for temporal comparison. Because digital shoreline position data are stored in a design file in the

order in which shorelines were compiled, sorting of points by location (UTM x- and y-coordinates) must be done upon export from the design file to ensure spatial consistency for temporal comparisons. Once the points are in order, the average orientation angle of the shoreline is used as a reference for calculating discrete positions for a user-defined interval. This step organizes randomly spaced information for a systematic comparison of spatial and temporal changes. Sample mean, standard deviation, 95 percent confidence limit, and percentage of shoreline represented by the statistics were tabulated for each cell.

Bathymetric Change

Hydrographic surveys of regional nearshore morphology provide a direct source of data for quantifying changes in sea-floor elevation. Historically, these data have been collected in conjunction with regional shoreline position surveys by the NOS. Comparison of digital bathymetric data for the same region but different time periods provides a method for calculating net movements of sediment into (accretion) and out of (erosion) an area of study. A number of manual and automated techniques have been used for making quantitative estimates of change. Moody (1964) superimposed contour data from charts of different time periods to identify change. Pierce (1969) used data point comparisons for exact geographic positions on charts of different time periods to calculate volumetric changes. Until the 1980's, these two procedures were standard practice (primarily contour overlay) for evaluating historical changes in nearshore bathymetry (e.g., Stauble and Warnke 1974), particularly those related to inlet systems (Dean and Walton 1975, Olsen 1977). However, in recent years, these data have been processed using commercially available surface modeling software (Hansen and Knowles 1988, List et al. 1991, Byrnes and Hiland, 1993).

Data Sources

Three main sources for bathymetric data were used for this study: Hydrographic maps (H-sheets) from NOS (formerly the US Coast and Geodetic Survey), digital bathymetric data from NOS, and digital bathymetric data from the USACE. NOS H-sheets represent the surveys performed in response to the act of February 10, 1807, establishing a national "Survey of the Coast" (Shalowitz 1964). Generally used for construction of navigation charts, these are considered the most accurate source of historical bathymetric data. Prior to the 1930's, surveys were performed using a graduated pole or a lead weight and graduated line. In the 1930's, the change was made to echo soundings, and later surveys used digital data collection methods. The hydrographic surveys were often preceded by a topographic survey to establish control points along the shoreline. Thus, T-sheets and H-sheets are usually available for similar dates, providing a useful and consistent data set for coastal zone studies. Bathymetric data from 1929, 1933, 1934, and 1962 were digitized from NOS H-sheets.

The National Geophysical Data Center (NGDC) in Boulder, Colorado maintains a digital database of the most recent NOS hydrographic surveys for US coastal areas. These data are delivered in ASCII files and contain information including sounding depths, longitude and latitude of soundings, the map number from which the sounding came, and the date of the map. For newer maps, these files contain all positions collected by the digital fathometer. The published maps are somewhat generalized and contain fewer points than the digital files. Maps produced prior to digital data collection are digitized at NGDC and contain all points on the published map. The 1977/78 data were imported from NGDC files to design files for comparison with other bathymetric data sets. In addition, a bathymetric survey was performed by the USACE-Baltimore District in June 1990 for the area covering the ebb tidal delta at Ocean City Inlet. The 1990 data was imported from DXF format files to design files for further processing.

Potential Errors

When comparing bathymetric data digitized from maps, the potential error sources for horizontal location of points is identical to those for comparison of shorelines. The data sets recorded digitally and imported directly (1977/78, 1990) contain less inherent errors due to the elimination of plotting positions onto a map and digitizing. While it is difficult to quantify the error in elevation due to errors in horizontal position, conservative estimates put this at between 0.3 and 0.5 meters, based on USC&GS and NOAA hydrographic manuals. Corrections to soundings for tides and sea level change introduce additional errors in vertical position of 0.1 to 0.3 meters. Finally, the accuracy of the depth measurement itself adds errors that are variable depending on the measurement method. For the 1929 and later surveys, all soundings have an accuracy of ± 0.3 meters (Shalowitz 1964).

Digitizing Bathymetric Data

The same set of guidelines presented for digitization of T-sheets applies to digitization of H-sheets. Scale, projection, and datum are taken into account when assessing each map. Files are created matching the cartographic parameters, and text strings are placed in the file in the proper position. Generally, the text on the map is centered on the point at which the sounding was taken. Therefore, the center point of the text on the map is digitized as the sounding location unless other indications of sounding location are present. For example, on some H-sheets, a dot is drawn at the sounding location. In this case, the location of the dot is digitized. After digitization, the files are converted to a common coordinate system.

The digital data from the NGDC is in an ASCII text format. Longitude and latitude locations, map numbers, dates, depths, and depth units are given for the sounding location. A program was written to read these data and place them into a file that is read by MicroStation. A map file was created in NAD 27 and text strings were then placed according to the locations given in the original file. This file then was converted to a coordinate system common to the other data sets.

Vertical Adjustments

Since ocean conditions are temporally and spatially inconsistent, adjustments to depth measurements must be made to bring all data to a common point of reference. These corrections include adjustments to compensate for tidal fluctuations for different surveys, changes in sea level over time, and differences in reference vertical datums.

All bathymetric data were corrected to the National Geodetic Vertical Datum of 1929 (NGVD). The 1990 depths were collected relative to Mean Low Water (MLW) at the Ocean City tide gauge, then corrected for tides and adjusted to the National Geodetic Vertical Datum of 1929 (NGVD). Changes in sea level were calculated for each previous survey according to the time elapsed and the rate of sea level change at their respective reference tide gauges (Byrnes et al. 1993). Also, the relationships between MLW and NGVD were obtained for each tide gauge. Because sea level has been rising and MLW is a lower elevation than NGVD, adjustments to all elevation values were applied with reference to NGVD.

Vertical adjustments were made after digitization of maps or importing of digital data, and after map files were brought to a common horizontal reference. Older maps often had values in feet, fathoms, and fractions of fathoms. Newer maps were in feet. An interactive program was written in the MicroStation Development Language to read in the digitized values, apply the appropriate vertical adjustment, and convert feet or fathoms to meters. The finest resolution of depths was approximately 0.1 m. Therefore, all depths and corrections were converted to meters, and all final values rounded to one decimal place.

Analysis of Bathymetric Data

Bathymetric data were analyzed using Intergraph's MGE Terrain Modeler, which operates within the framework of the Modular GIS Environment. This means that simultaneous access to shorelines, engineering structures, bathymetry, annotation, and other data is possible. The interactive nature of the software allows incorporation of shoreline and structure information into the established terrain model, providing an accurate and complete representation of the seafloor and adjacent boundaries. Polygon boundaries were defined for the study area by bathymetric data limits. The 1929/33, 1961/62, and 1977/78 data covered a region stretching from approximately 5 km north to 5 km south of Ocean City Inlet, and approximately 3 km offshore. The 1990 data only covered the ebb-tidal delta area at the inlet. Therefore, a subset of the other data sets was used for comparison with the 1990 data. The landward boundary condition was determined by overlaying all digitized shoreline position data and creating a composite edge of the most landward positions. This method insures that all areas that show change over the period of record are included in the analysis.

Unlike shoreline change analysis where horizontal position is the only variable necessary to quantify change, terrain modeling requires a vertical measurement at each point to create a model

surface for temporal and spatial comparisons. Because the landward boundary was defined as the position of a composite landward high-water shoreline, an elevation for this position was estimated using available wave data (Jensen 1993) and beach profile information. The position monitored by topographers and photo-interpreters during shoreline surveys is defined visually as the line associated with the average wave uprush during high tide (Shalowitz 1964). The elevation of the HWL was estimated to be 0.9 meters above NGVD for the Ocean City area (see Byrnes and Hiland, 1994 for details of this procedure).

Digitized soundings, shorelines, structures, and boundary polygons were used to create digital terrain models of the seafloor for each time period. The Triangulated Irregular Network (TIN) was the terrain modeling routine used in this study to form a sheet of continuous connected triangular planes based on irregular points. The elevation of each point is determined by solving equations for its horizontal location on the triangulated surface. Therefore, only points existing in the original data sources are used to create the surface model, as opposed to grid models which interpolate evenly spaced points from original data. TIN models were used for all calculations of bathymetric change; however, grid surfaces were generated for graphic display purposes.

To calculate volumes of change, TIN models representing two different dates were compared using the Compute Volume function within MGE Terrain Modeler software. Primary and secondary surfaces are declared prior to change calculation. Every node from the primary surface is projected onto the secondary surface and the z value of the secondary surface is subtracted from the z value of the original point. Likewise, every node in the secondary surface is projected onto the primary surface and the z values subtracted. The resultant TIN model then contains a point corresponding to each point in the primary and secondary surfaces, with the z value representing the differences in elevation between the two. Volumes of change are calculated by summing the volume of each triangle region.

6 Results

Shoreline Change

Calculation of shoreline change rates encompass the area approximately 19 km south of Ocean City Inlet (Assateague Island) to approximately 6 km north of Ocean City Inlet (Fenwick Island). Shoreline data covering the last 140 years show a pronounced concave shoreline configuration, produced by the interruption of longshore sediment transport (Figure 44). Historical shoreline change rates between 1850 and 1989 for Fenwick Island averaged -0.4 m/yr, and -1.9 m/yr for the Assateague portion of the study area. However, spatial and temporal differences in shoreline position change rates are highly variable in this area, as illustrated in Figure 45. To provide a more accurate representation of shoreline changes, the Assateague Island shoreline was divided into two zones. Zone A extends from the inlet south 6.5 km, and Zone B extends from 6.5 km - 19 km. The shoreline change rates were divided into two time periods (pre and post inlet conditions). The first period 1850-1933, is representative of pre-inlet shoreline change rates. The second time period, 1933-1989 represents post-inlet shoreline change rates. Average rates of shoreline position change for all combinations of time periods, for both Fenwick and Assateague, are shown in Tables 6 and 7.

Pre-Inlet Conditions 1850-1933

Fenwick Island

Shoreline change rate data for the period 1850, 1908, and 1933 represent conditions prior to the opening of Ocean City Inlet. As stated earlier in the text, the inlet breached adjacent to an existing wooden groin. The average shoreline change rate from 1850-1908 for the area north of this groin was -1.5 m/yr. Between 1908 and 1933, the retreat rate decreased to -1.2 m/yr, probably due to updrift sediment accumulation resulting from various groin features constructed during this time period. Consequently, the average rate of shoreline retreat for the 83 year time period was -1.4 m/yr. Cumulative shoreline change rates from 1850-1933 are illustrated in Figure 46.

Assateague Island

South of the wooden groin where the inlet breached, Assateague Island's average shoreline change rate from 1850-1908 was -2.70 m/yr (zone A) and .84 m/yr (zone B). The shoreline change rate increased to -4.23 m/yr (zone A) and -1.99 m/yr (zone B) from 1908-1933. Some percentage of this increased rate of shoreline retreat may be a result of the downdrift erosional *shadow* that typically results after construction of groins. These data may give some indication of Assateague Island's vulnerability to breaching, and may indicate the presence of man-induced

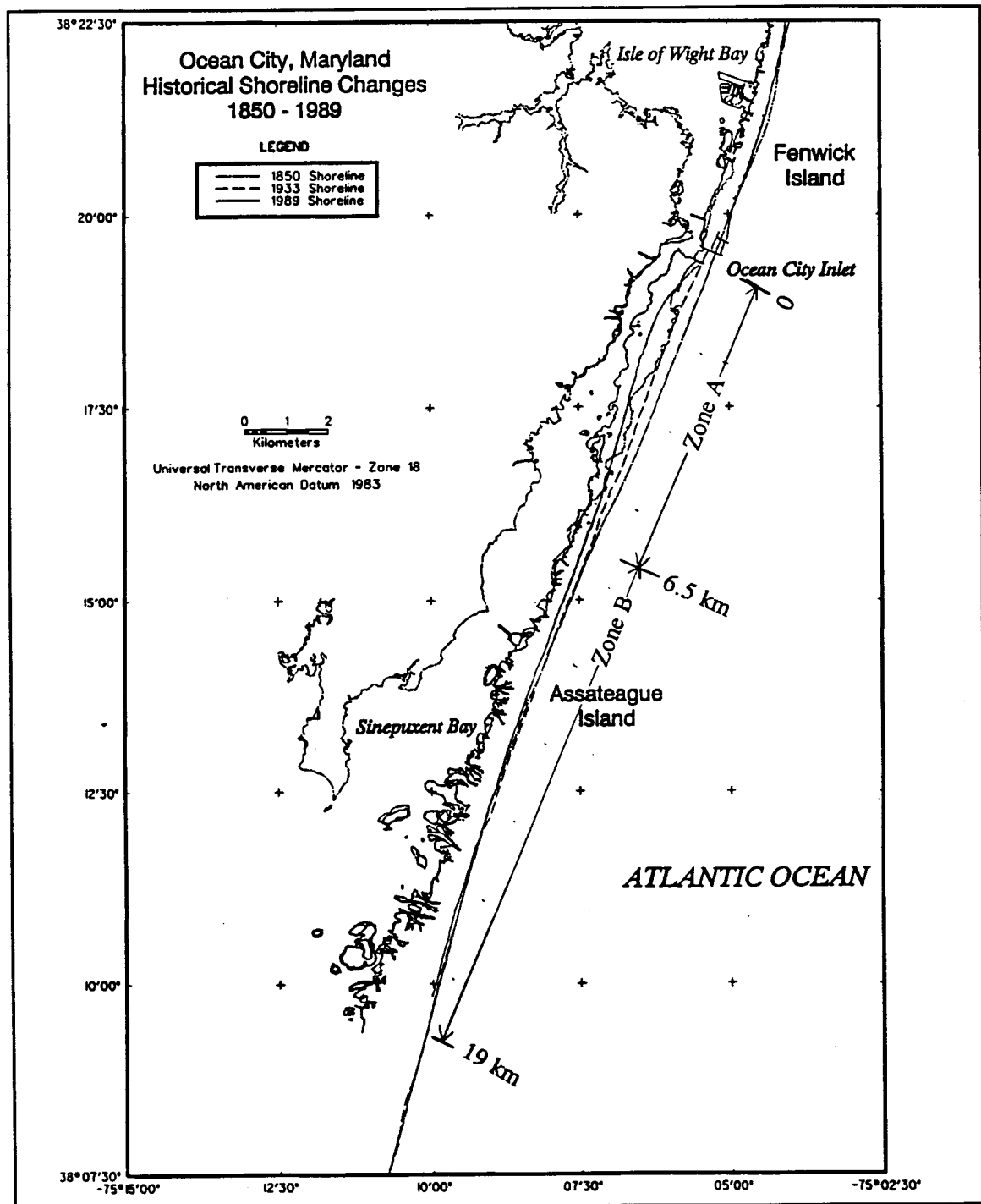


Figure 44. Historical shoreline changes for Fenwick and Assateague Island from 1849-1933-1989. Note concaved shoreline configuration adjacent to the inlet.

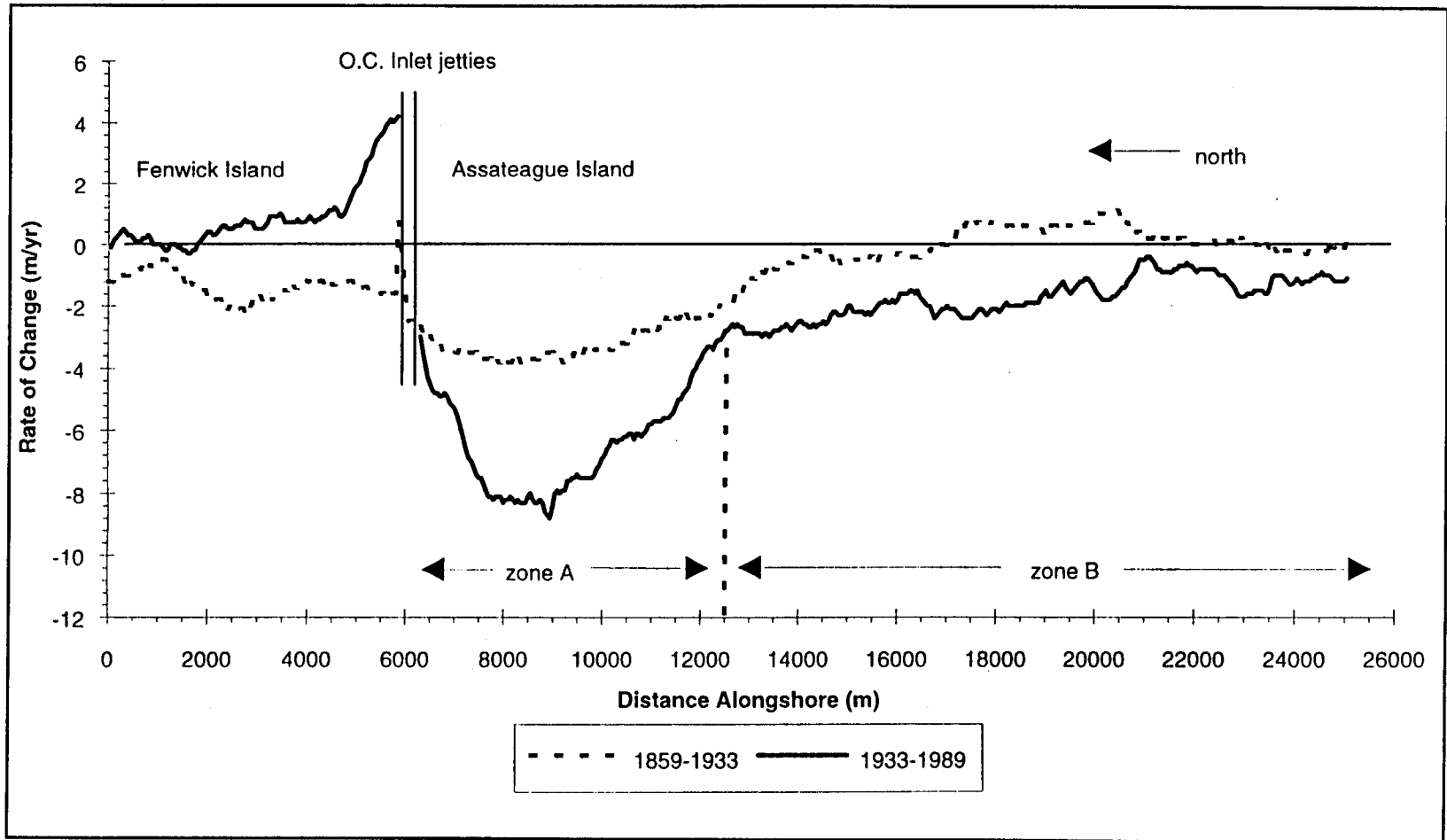


Figure 45. Shoreline position changes (m/yr) along the 25 km study area from 1850-1933 and 1933-1989. Note the high variability adjacent to Ocean City Inlet in zone A as compared to zone B.

Table 6. Average Shoreline Position Change (m/yr) for the Southern 5.8 km of Fenwick Island, Pre and Post Inlet Conditions

Year	1908 ⁴	1933	1942	1962	1976	1980	1989
Pre-inlet 1850	-1.5 ¹	-1.4	-1.0	-0.6	-0.5	-0.5	-0.4
	0.4 ²	0.4	0.3	0.6	0.6	0.5	0.5
	0.1 ³	0.1	0.1	0.1	0.1	0.1	0.1
1908	----	-1.2	0.2	0.2	1.1	0.8	0.9
	----	0.5	0.6	1.1	1.2	0.9	0.9
	----	0.1	0.2	0.3	0.3	0.3	0.2
1933	----	----	1.2	1.2	1.0	0.8	0.9
	----	----	1.9	1.9	1.6	1.2	1.1
	----	----	0.3	0.4	0.3	0.2	0.2
Post-inlet 1942	----	----	----	1.3	1.0	0.6	0.9
	----	----	----	2.4	1.8	1.2	1.0
	----	----	----	0.4	0.3	0.2	0.2
1962	----	----	----	----	0.4	-0.2	-0.3
	----	----	----	----	1.1	0.8	0.6
	----	----	----	----	0.2	0.1	0.1
1976	----	----	----	----	----	-2.1	0.5
	----	----	----	----	----	3.7	1.2
	----	----	----	----	----	0.7	0.2
1980	----	----	----	----	----	----	1.7
	----	----	----	----	----	----	1.5
	----	----	----	----	----	----	0.3

¹average shoreline change rate (m/yr); ²standard deviation ($\pm$ m/yr); ³95 percent confidence limit ($m \pm yr$); ⁴ data extends only 3.85 km south

Table 7. Average Shoreline Position Change (m/yr) for the Northern 19 km of Assateague Island, Pre and Post Inlet Conditions.

Year	1908		1933		1942		1962		1976 ⁴	1980		1989	
	A	B	A	B	A	B	A	B	A	A	B	A	B
Pre-inlet 1850	-2.70 ¹	.84	-3.15	-0.01	-3.49	0.10	-4.50	-0.20	-5.4	-4.58	-0.18	-4.44	-0.70
	1.06 ²	0.59	0.53	0.58	0.92	0.45	1.44	0.41	0.48	1.10	0.53	0.95	0.52
	0.18 ³	0.07	0.09	0.07	0.16	0.05	0.25	0.05	0.10	0.19	0.06	0.16	0.06
	-----	-----	-4.23	-1.99	-4.83	-1.14	-6.51	-1.36	-7.18	-6.09	-1.01	-5.68	-1.81
1908	-----	-----	1.16	-1.32	0.81	0.82	1.91	0.67	0.65	1.22	0.85	1.14	0.77
	-----	-----	0.20	0.16	0.14	0.10	0.34	0.08	0.14	0.21	0.10	0.20	0.09
	-----	-----	-----	-----	-6.37	1.22	-8.41	-0.81	-9.16	-7.06	-0.47	-6.30	-1.72
1933	-----	-----	-----	-----	5.11	1.59	4.31	0.49	1.06	2.18	0.75	1.65	0.66
	-----	-----	-----	-----	0.89	0.19	0.77	0.06	0.23	0.38	0.09	0.29	0.08
	-----	-----	-----	-----	-----	-----	-9.41	-1.7	-9.04	-7.23	-0.88	-6.28	-2.30
Post-inlet 1942	-----	-----	-----	-----	-----	-----	2.43	4.22	1.15	1.68	1.07	1.63	0.87
	-----	-----	-----	-----	-----	-----	0.44	0.76	0.25	0.29	0.13	0.28	0.10
	-----	-----	-----	-----	-----	-----	-----	-----	-4.42	-4.52	0.06	-3.78	-2.71
1962	-----	-----	-----	-----	-----	-----	-----	-----	3.83	3.08	1.57	3.92	1.14
	-----	-----	-----	-----	-----	-----	-----	-----	0.89	0.55	0.19	0.70	0.14
	-----	-----	-----	-----	-----	-----	-----	-----	-----	-1.57	-----	-0.13	-----
1976	-----	-----	-----	-----	-----	-----	-----	-----	-----	5.99	-----	6.03	-----
	-----	-----	-----	-----	-----	-----	-----	-----	-----	1.33	-----	1.35	-----
	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-2.30	-8.20
1980	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	6.34	1.32
	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	1.11	0.16
	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----

¹average shoreline change rate (m/yr); ²sample standard deviation ($\pm$ m/yr); ³95 percent confidence limit ($\pm$ m/yr); ⁴data extends only 3.85 km south.

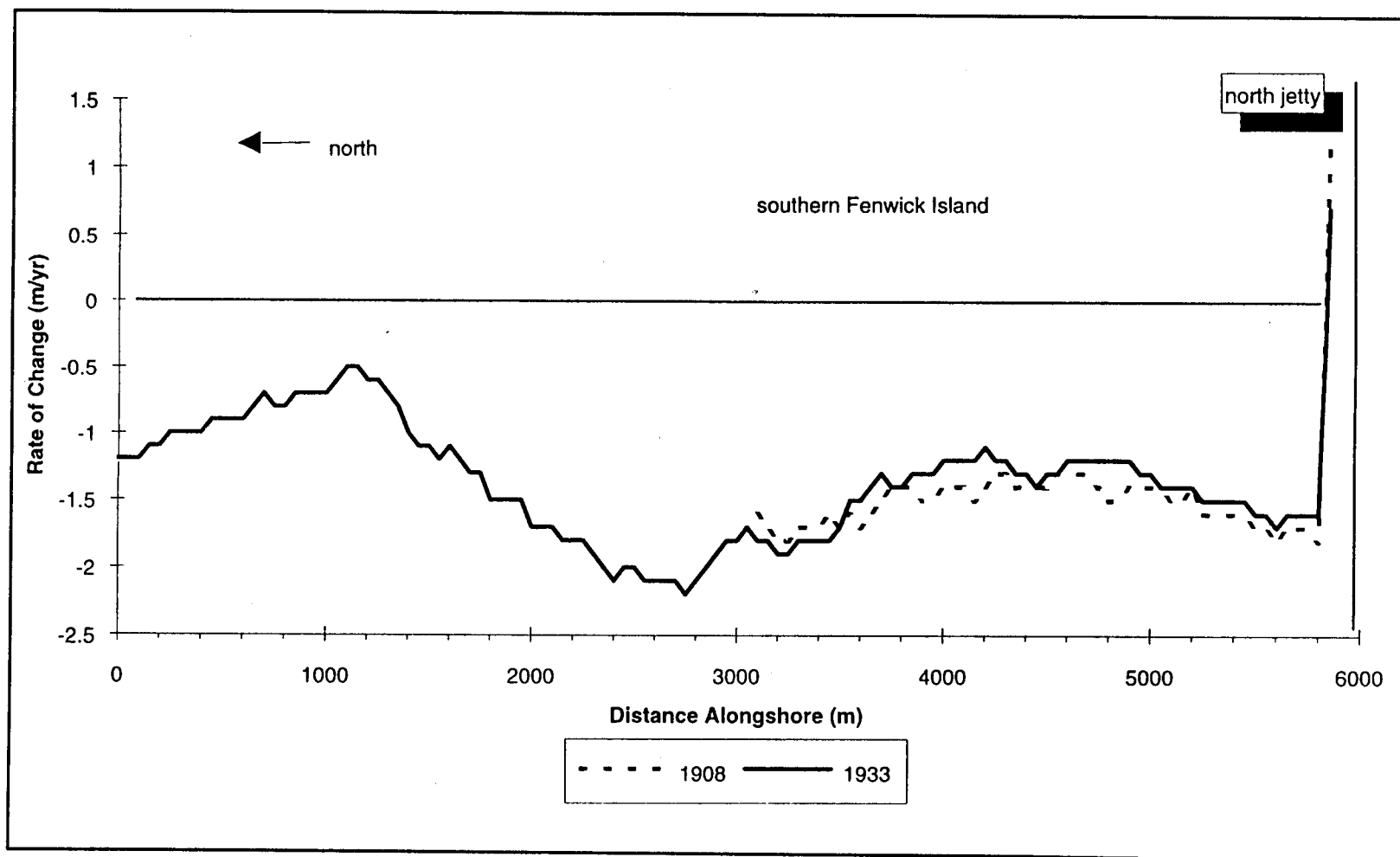


Figure 46. Cumulative shoreline change rates from 1850-1933 for southern Fenwick Island, pre-inlet conditions.

erosion. Figure 47 depicts cumulative rates of shoreline position change for northern Assateague Island from 1850-1933.

Post-Inlet Conditions 1933-1989

Fenwick Island

Shoreline change rate data for the periods 1942, 1962, 1976, 1980, and 1989 represent conditions after jetty stabilization at Ocean City Inlet. The average shoreline change rate reversed from -1.2 m/yr in 1933 to 1.2 m/yr in 1942. This shoreline advancement adjacent to the north jetty, produced by the interruption of longshore sediment transport, is depicted in (Figure 48). Since 1942, there has been little if any significant change in shoreline advancement adjacent to the north jetty. One reason for this may be because the north fillet adjacent to the jetty is impounded to capacity. Sediment arriving near this region would be moved along and around the seaward end of the north jetty. Dean and Perlin (1977) state; "the north jetty is impounded to capacity" and that "in the period after 1934-1947, an average of 76 percent of the net material arriving at the north jetty would be bypassed". This 6 km stretch of shoreline has assumed more of an equilibrium configuration in response to jetty construction over this 56 year time period (Figure 49).

Assateague Island

Shoreline change rates from 1933 through 1942 are shown in Figure 50. Between 1933-1942, an area of severe shoreline retreat developed just downdrift of the south jetty, indicating an interruption of longshore sediment transport processes. Over this 9 year time span, average shoreline retreat rates for zone A increased from -3.15 m/yr to -6.37 m/yr. In zone B, shoreline change rates averaged 1.22 m/yr.

From 1942 to 1962, zone A retreat rates increased to -9.41 m/yr. The 1962 shoreline was mapped after the "Ash Wednesday" storm of March 1962. The effects of this storm are evident on the shoreline change map (Figure 51). A storm induced breach along the south jetty, and a breach (approximately 2000 m) downdrift of the south jetty, imply a vulnerability of this shoreline to breaching due to sediment deprivation to northern Assateague Island. Cumulatively, the average rate of shoreline retreat from 1933-1962 was -8.41 m. Zone A standard deviation values for 1942 and 1962 averaged approximately ± 5.1 m/yr and ± 4.3 m/yr, respectively. This value is a measure of sample variability, indicating a very irregular or large fluctuation of shoreline change rates, indicative of dynamic changes within this section of shoreline.

Shoreline change rates from 1962 to 1976 are shown in Figure 52. Between 1962 and 1976 average shoreline retreat rates for zone A was approximately -4.42 m/yr. This reduced shoreline change rate may be attributable to emergency USACE repairs to this stretch of shoreline. Approximately 760,000 m³ of dredged material was used to reestablish the shoreline. The reduced

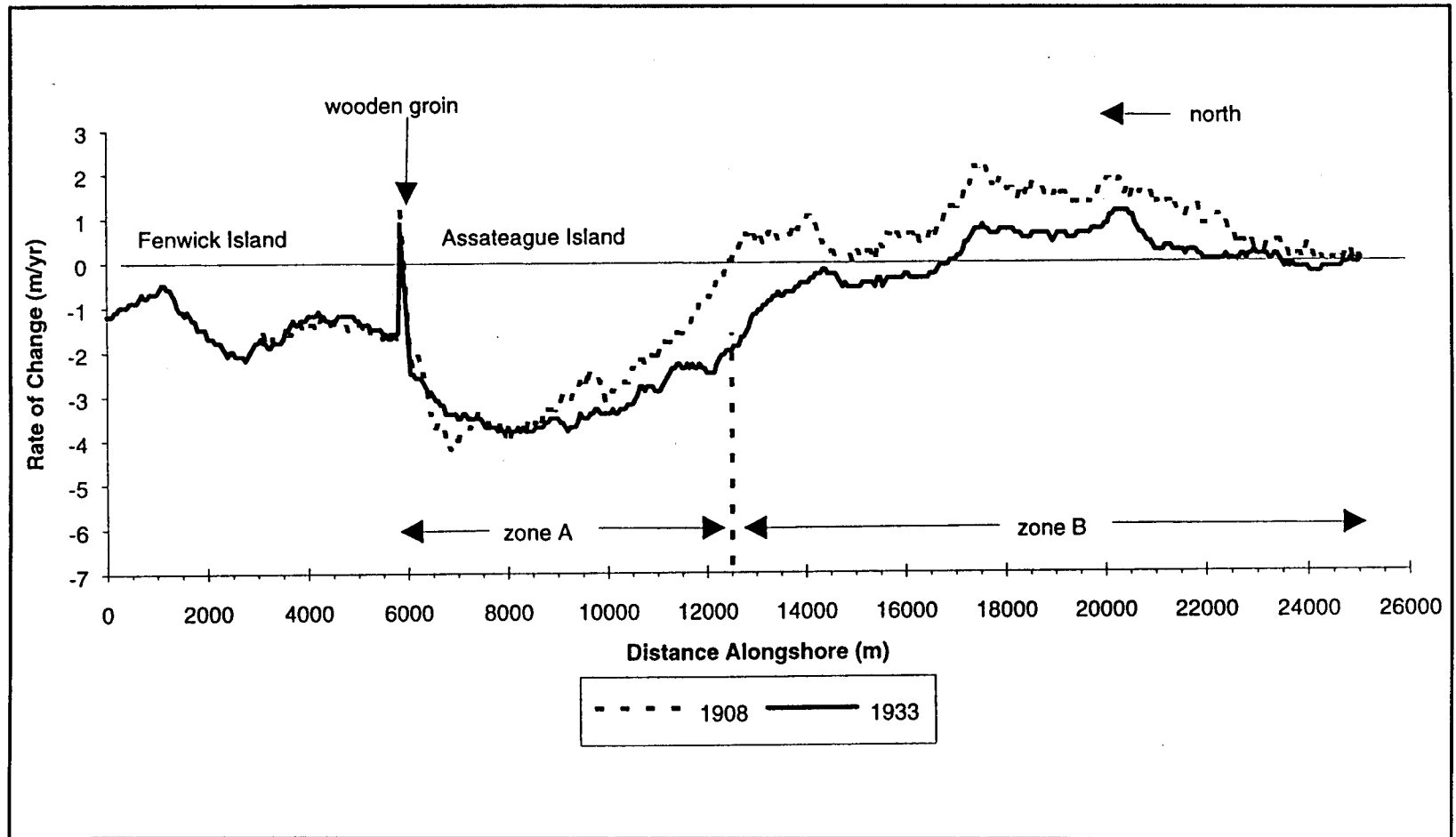


Figure 47. Cumulative rates of shoreline position change for northern Assateague Island, Maryland from 1850-1933, pre-inlet conditions. The inlet breached adjacent to a wooden groin.

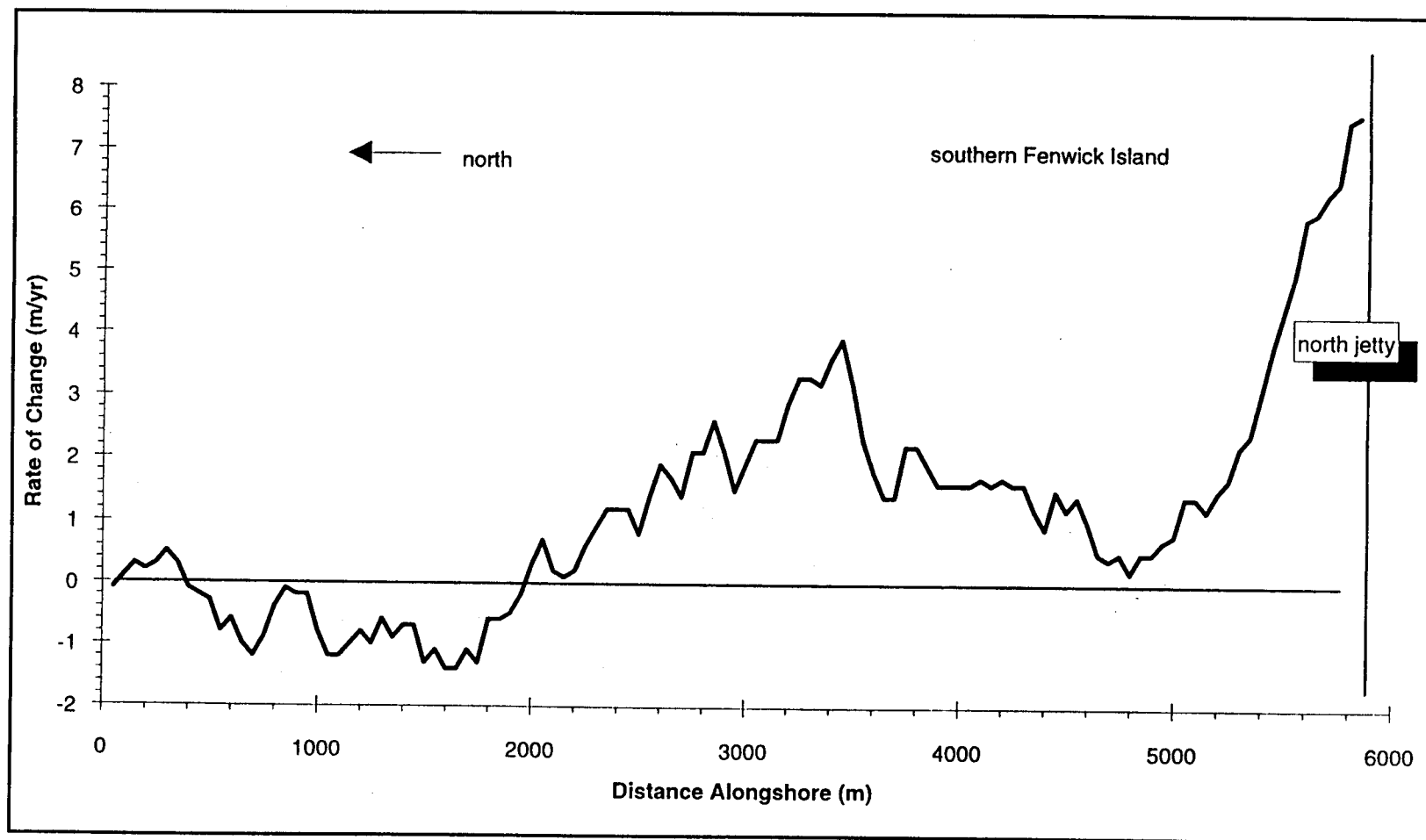


Figure 48. Shoreline advancement adjacent to the north jetty produced by the interruption of longshore sediment transport since inlet stabilization in 1933.

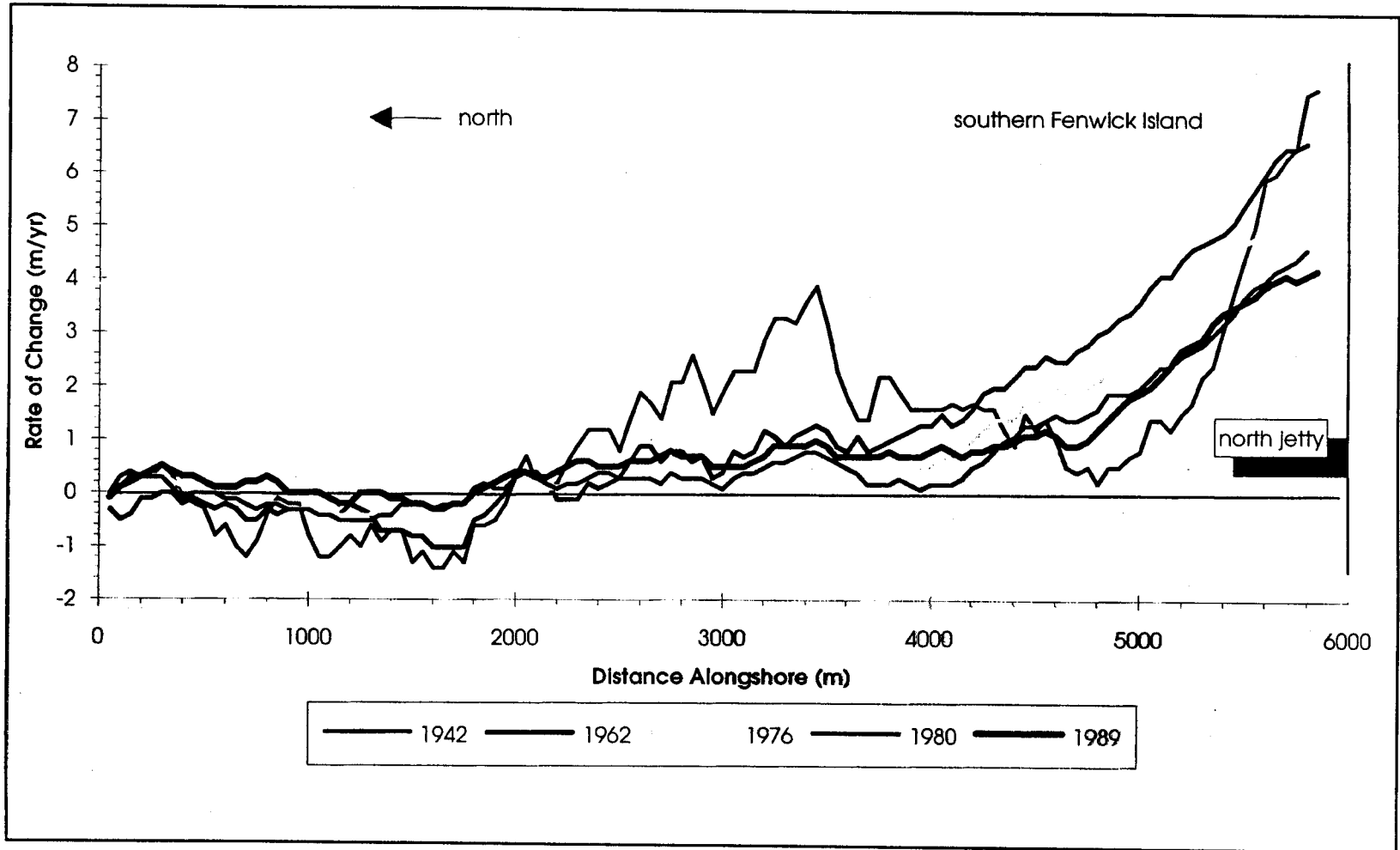


Figure 49. Cumulative rates of shoreline position change for southern Fenwick Island, Maryland from 1933-1989, post inlet conditions.

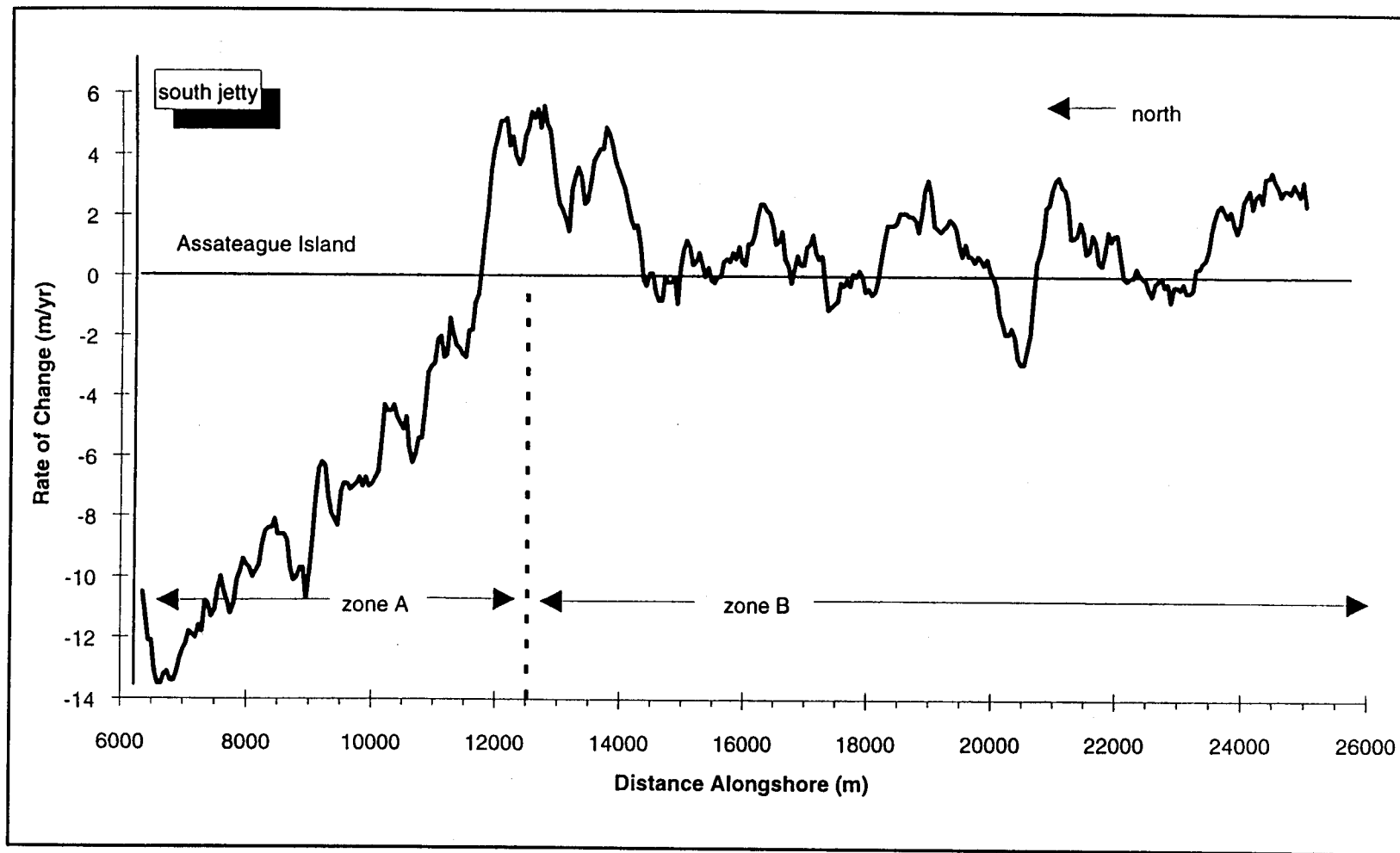


Figure 50. Short term rate of shoreline position change for northern Assateague Island, Maryland from 1933-1942.

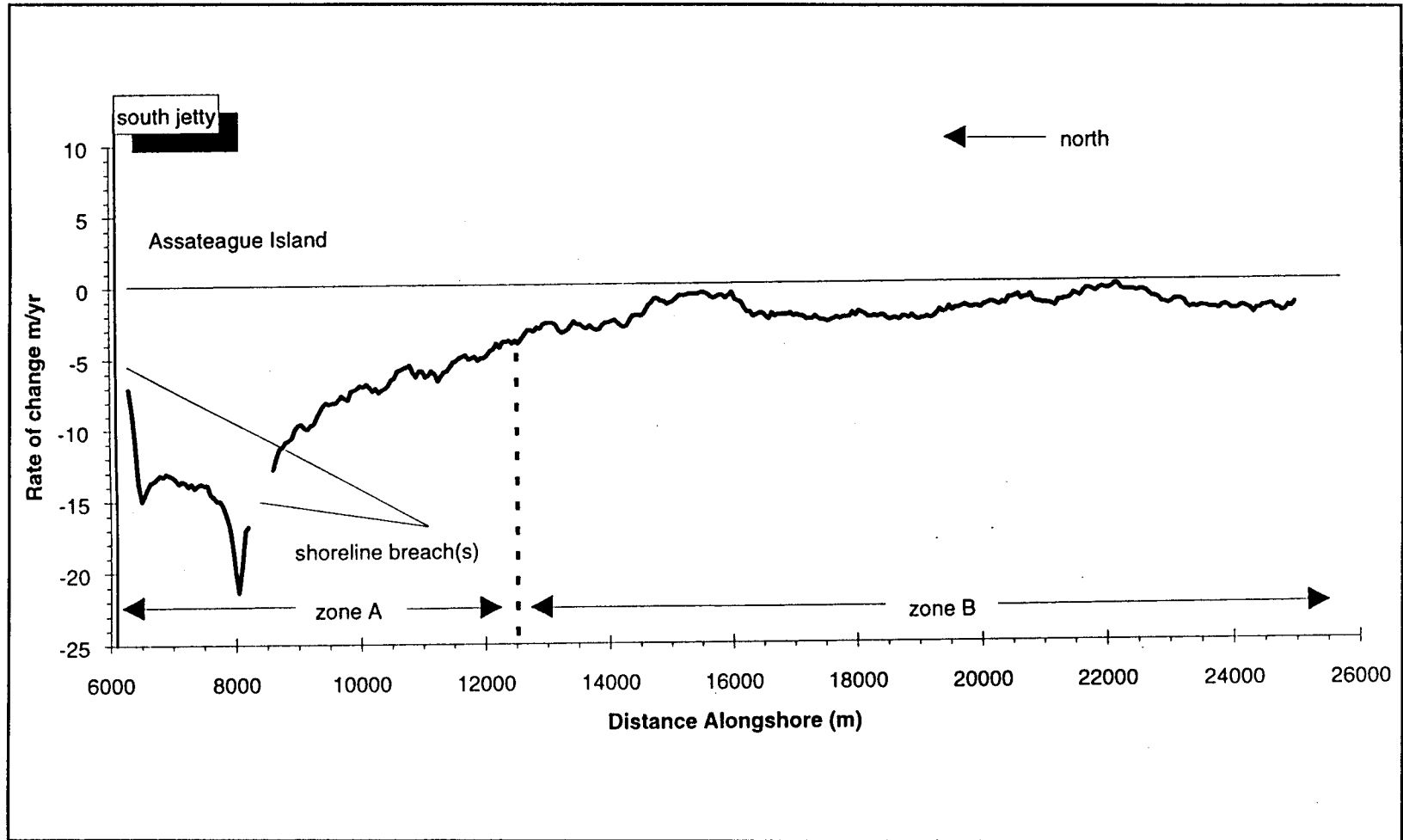


Figure 51. Short term rate of shoreline position change for northern Assateague Island, Maryland from 1942-1962.

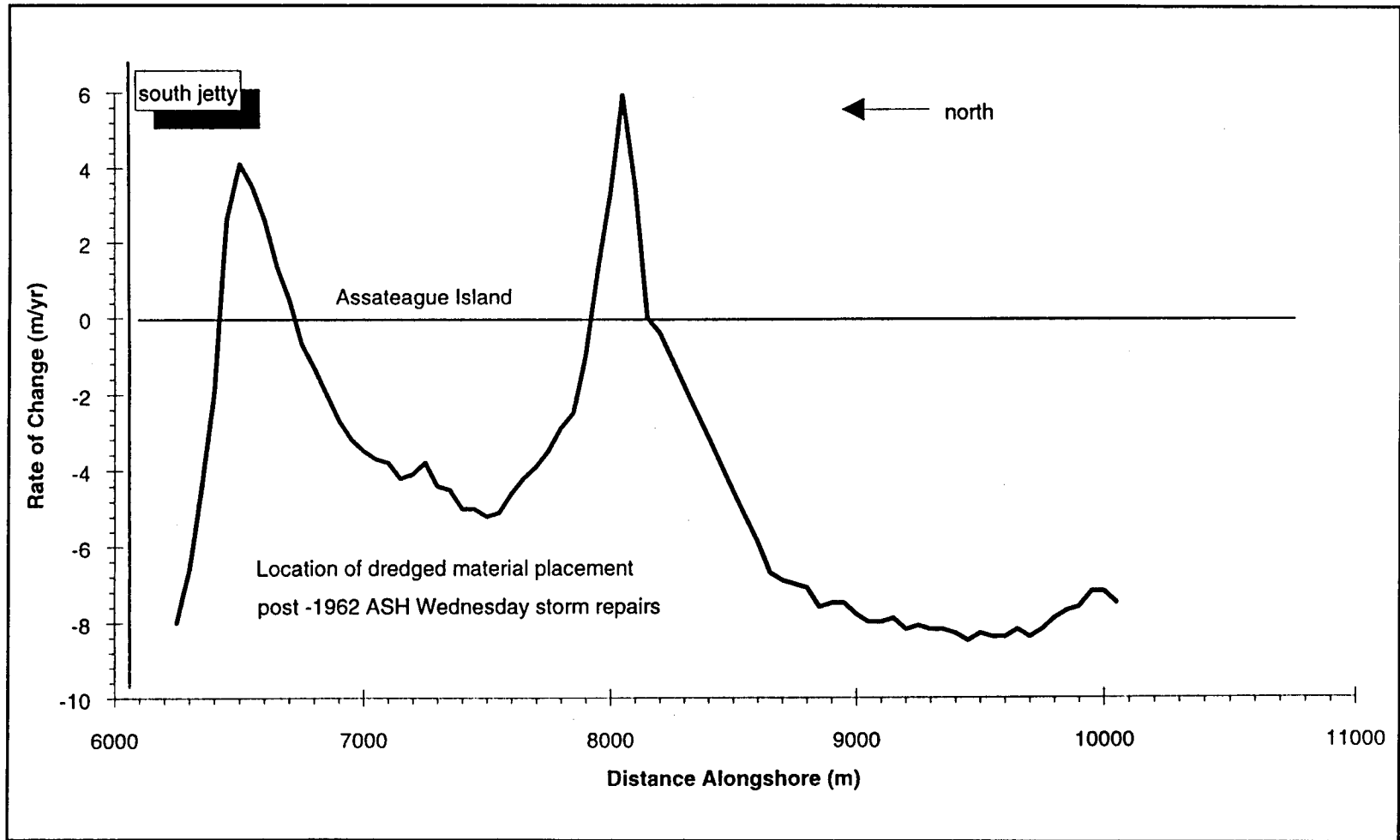


Figure 52. Short term rate of shoreline position change for northern Assateague Island, Maryland from 1962-1976.

rates may also be due to only having 3.85 km of shoreline data for this time period. However, the longer time period of 1933-1976 still indicates shoreline change rates on the order of -9.0 m/yr.

Shoreline change rates from 1976 through 1980, are shown in Figure 53 for zone A due to limited data. This time period may mark the beginning of actual sediment bypassing to northern Assateague Island. Shoreline change rates from 1976 to 1980, are very erratic ranging +15 m/yr to -10 m/yr. The standard deviation associated with this value was ± 6 m/yr, indicating a high variability in changes along this 3.85 km segment of shoreline. Cumulative average shoreline retreat rates from 1933-1980 for zone A were -7.06 m/yr. This longer length of time (67 years) clearly shows reduced average shoreline retreat rates, reversing the trend of increasing shoreline retreat since 1933.

Figure 54, depicts shoreline change rates from 1980-1989. The shoreline change rate for zone A has decreased from -7.06 m/yr for 1933-1980 to -6.30 m/yr for 1933-1989. This is due to an area of accretion that developed adjacent to the south jetty, beginning around the 1976/1980 time period and extending through 1989. Between 1980-1989 zone B shows an average shoreline change rate of -8.2 m/yr. This shoreline retreat for zone B seems uncharacteristically high considering historical changes for this zone since 1850. However analysis of offshore wave data by Bass et al (1994) show approximately 13 storm events from August 1986 to December 1987. Wave heights associated with these storms ranged from 2 m to 5 m. Dolan et al (1990) reported the effects of another strong northeaster in March 1989 that heavily damaged the Virginia and North Carolina shorelines. From March 7-11, significant wave heights exceeded 1.5 m over a continuous 115 hour time span. Since the aerial photography was taken on April 12, the higher shoreline retreat rates for zone B may be indicative of these repetitive storm events. Between 1933-1980 and 1933-1989, zone B averaged -0.47 m/yr and -1.72 m/yr, respectively.

Since 1976, shoreline retreat rates have been steadily reduced from the inlet south for approximately 4 km (Figure 55). Between 4 km and 6 km, virtually no reduction in retreat rates have occurred, and from 6 km south, retreat rates are similar to pre-inlet conditions. Figure 56 shows the location adjacent to the south jetty where the shoreline has advanced since 1976. This shoreline segment is approximately 1.6 km in length, and encompasses an area of 133,760 m². Assuming that .09 m² of beach change is equivalent to about .76 m³ of sand, results in an overall estimated accretion volume of approximately 1.13 million m³ of sand (March 1976-April 1989).

This produces an overall estimated accretion of approximately 87,000 m³/yr. This annual accretion rate is similar to results found by Bass et al. (1994). Their volume estimates from beach profile data (August 1986 to March 1990) from the inlet south 1.2 km, estimate a accretion volume rate of approximately 121,600 m³/yr.

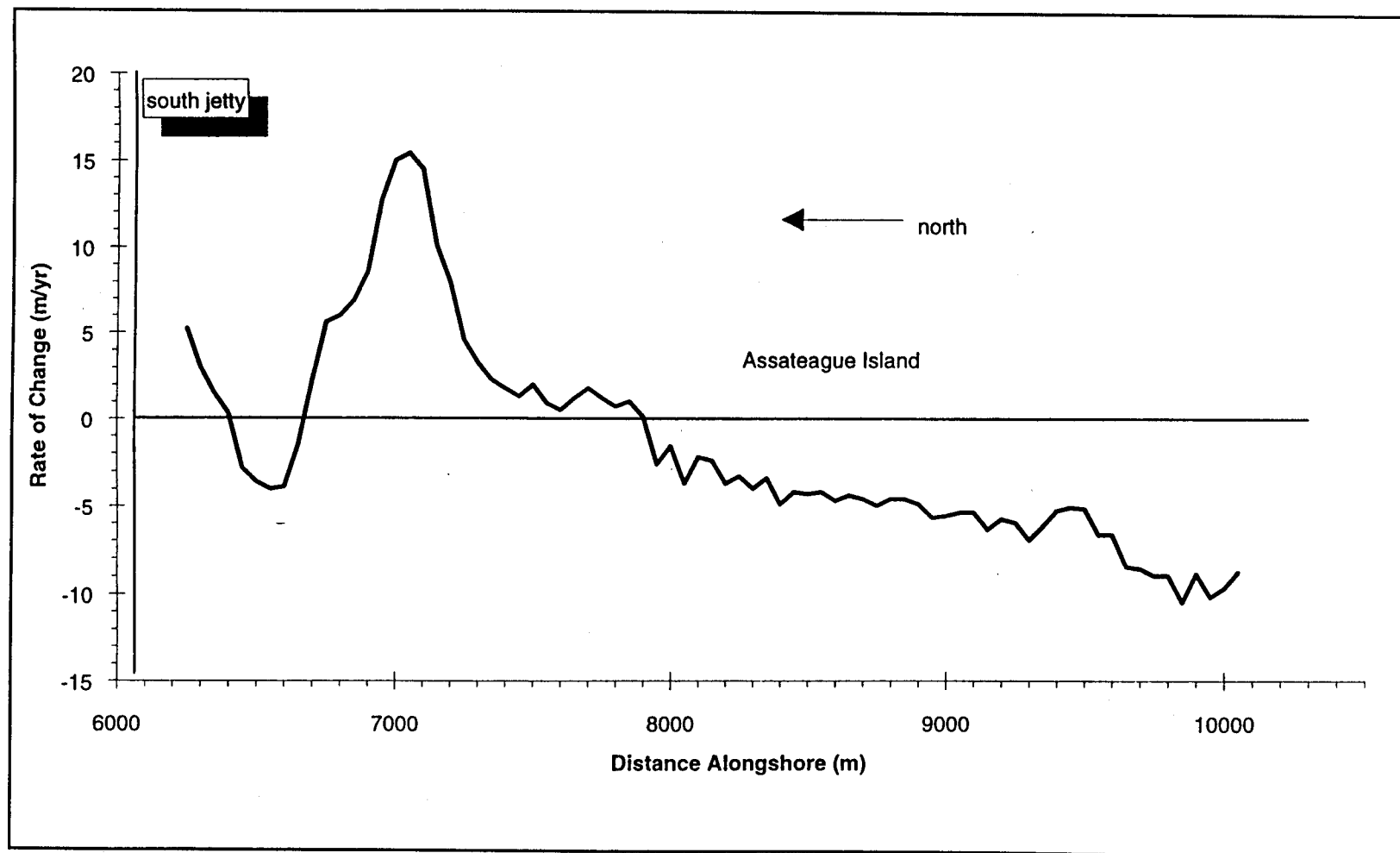


Figure 53. Short term rate of shoreline position change for northern Assateague Island, Maryland from 1976-1980.

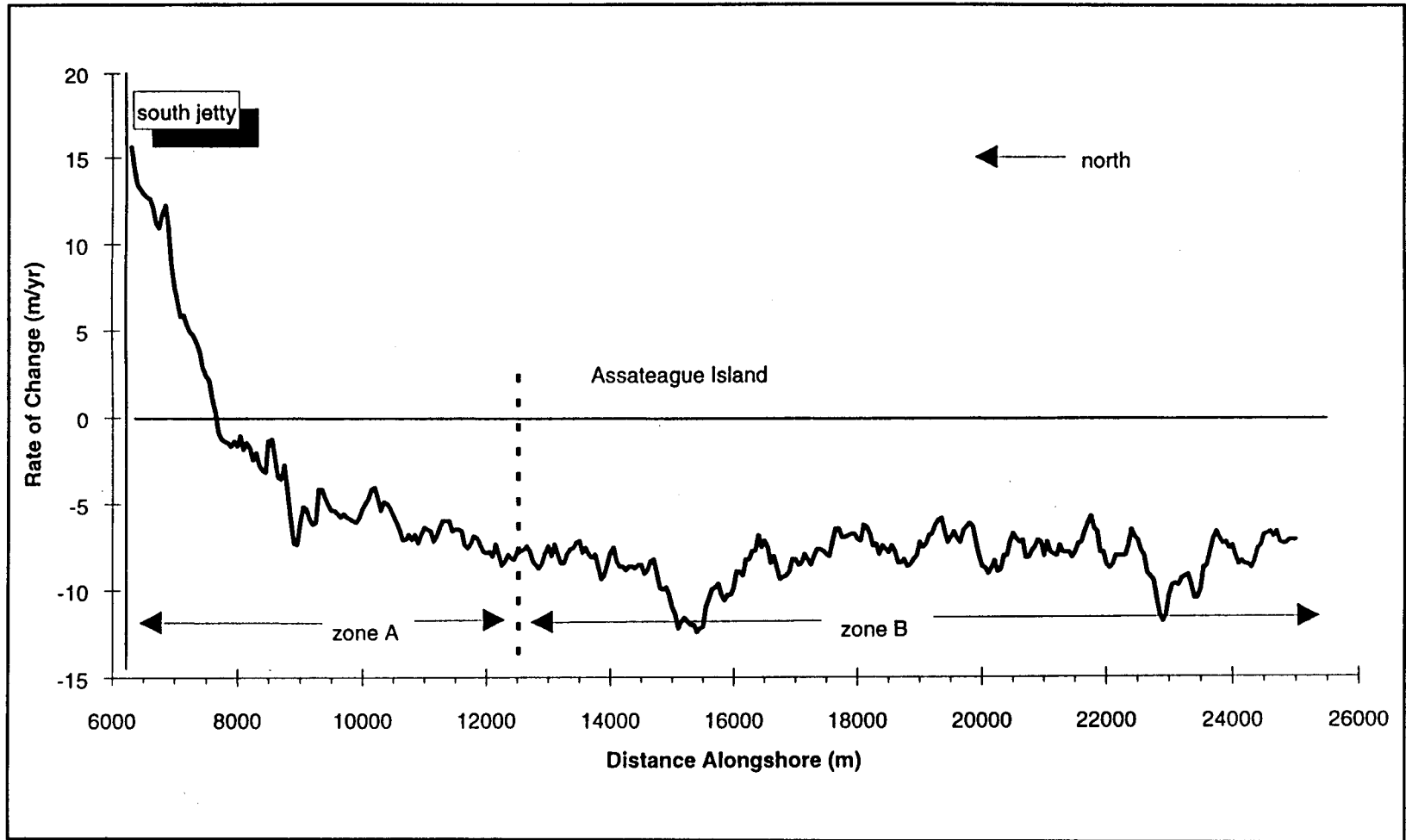


Figure 54. Short term rate of shoreline position change for northern Assateague Island, Maryland from 1980-1989.

Ebb Tidal Delta Growth

Four time periods were used to quantify ebb shoal growth. The 1933 data set is representative of the pre-inlet bathymetric conditions. The 1962, 1978, and 1990 data represent post inlet bathymetric conditions. In order to eliminate any seasonal differences, all bathymetry data used for this study fell within the same spring-summer collection period. After review of all the bathymetric data, the -7 m contour was used to identify the outer edge or terminal lobe of the ebb shoal. Identification of this contour allowed for measurements of offshore growth and lateral width, in comparison with the location of the -7 m contour in 1933.

The nearshore bathymetry in June 1933 is shown in figure 57. Prior to inlet formation, the nearshore bathymetry consisted of straight parallel contours from -2 m to -8 m. The irregular bathymetric influence of an offshore linear shoal complex is evident along the -10 m contour.

Nearshore bathymetric changes from 1933-1962, represented by movement of the 7 m contour, are shown in figure 58. This figure shows the ebb shoal nearly overtopping an existing linear shoal that runs northeast-southwest. Over this 29 year period, the ebb shoal's outer portion extended seaward to a distance of 1.1 km offshore, and 1.3 km in width. The shoal has assumed a crescentic shape, and a downdrift offset configuration towards northern Assateague Island (Figure 59). The shallower portions of the shoal platform are noticeable along the southeastern portion, with vertical changes of 5 m within the interior, and 2 m along the outer edge (Figure 60).

Over the next 16 year period (1962-1978), the ebb shoal advanced seaward only 70 m, while its width increased approximately 324 m (Figure 61). In figure 62, the 1978 bathymetry is depicted with a more pronounced extension of the ebb shoal moving toward attachment to the downdrift shoreline of northern Assateague Island. Vertical shoal growth was primarily confined to the southeastern portion (increasing by 2 m), and smaller segments along the eastern and northern flanks (Figure 63).

From 1978-1990, the ebb shoal advanced seaward approximately 100 m offshore, and increased 240 m in width (Figure 64). The 1990 bathymetry depicts a well defined shoal platform and outer boundaries (Figures 65). Figure 66 shows attachment of the shoal to northern Assateague Island, with 2 m vertical increases in elevation along the entire outer portion of the shoal. Another interesting note from this figure is the contribution of the linear shoal complex to the accretion of the ebb shoal along the 2 m contour. In figure 67, two adjacent bar ridges have formed along the southeastern edge, possibly evidence of continued lateral growth and the presence of the linear shoal complex. From this 1990 bathymetry, Ocean City inlet ebb shoal extends offshore from the jetties approximately 1.3 km offshore, and is 1.9 km wide. The shoal is crescentic shaped, offset to the south and attached directly to the northern end of Assateague Island, approximately 600 m south of the south jetty.

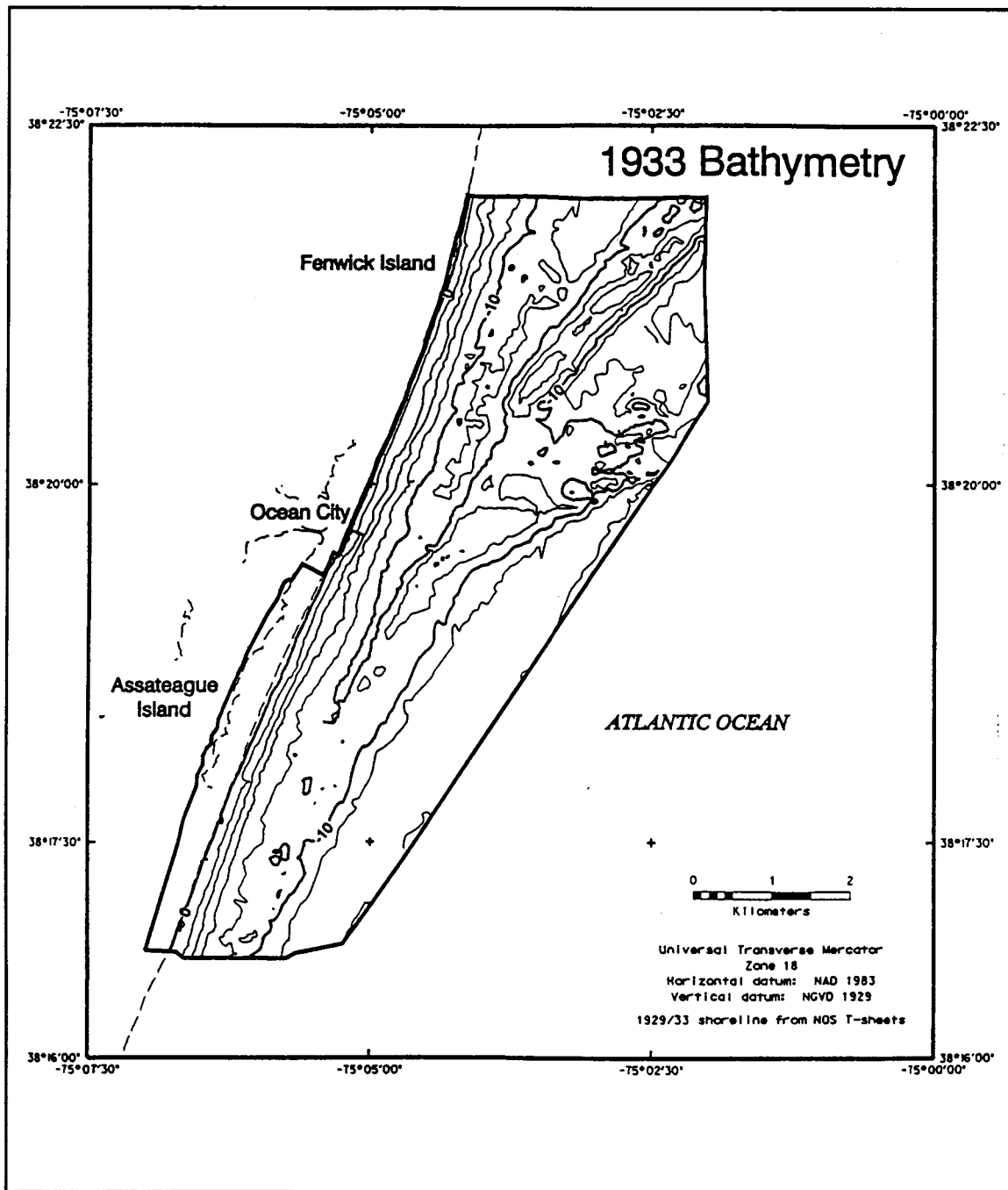


Figure 57. 1933 bathymetric contours prior to inlet breaching. Note straight and level contours adjacent to Ocean City Inlet.

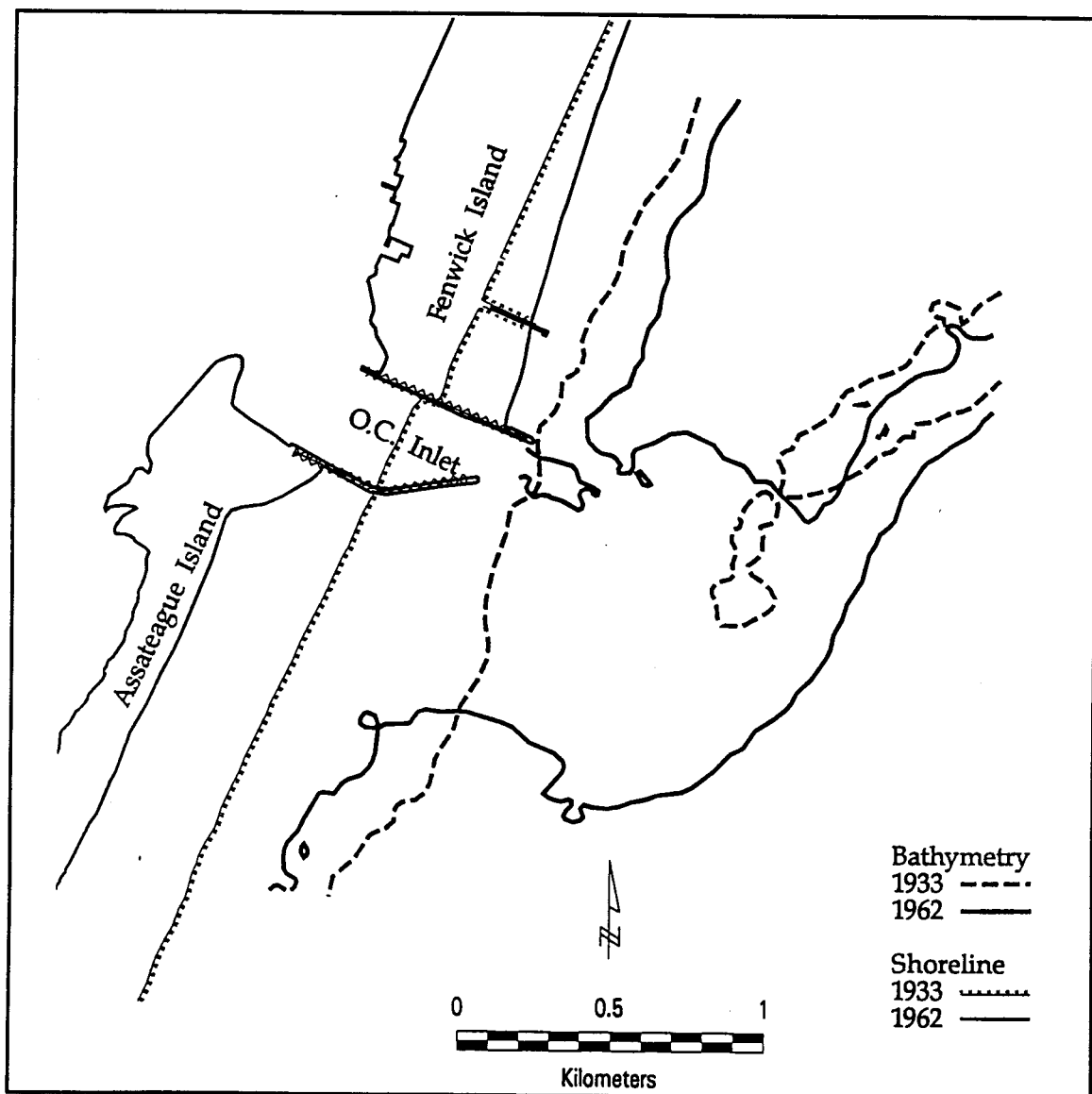


Figure 58. -7 m contour changes from 1933-1962. Note the offshore movement across a linear shoal.

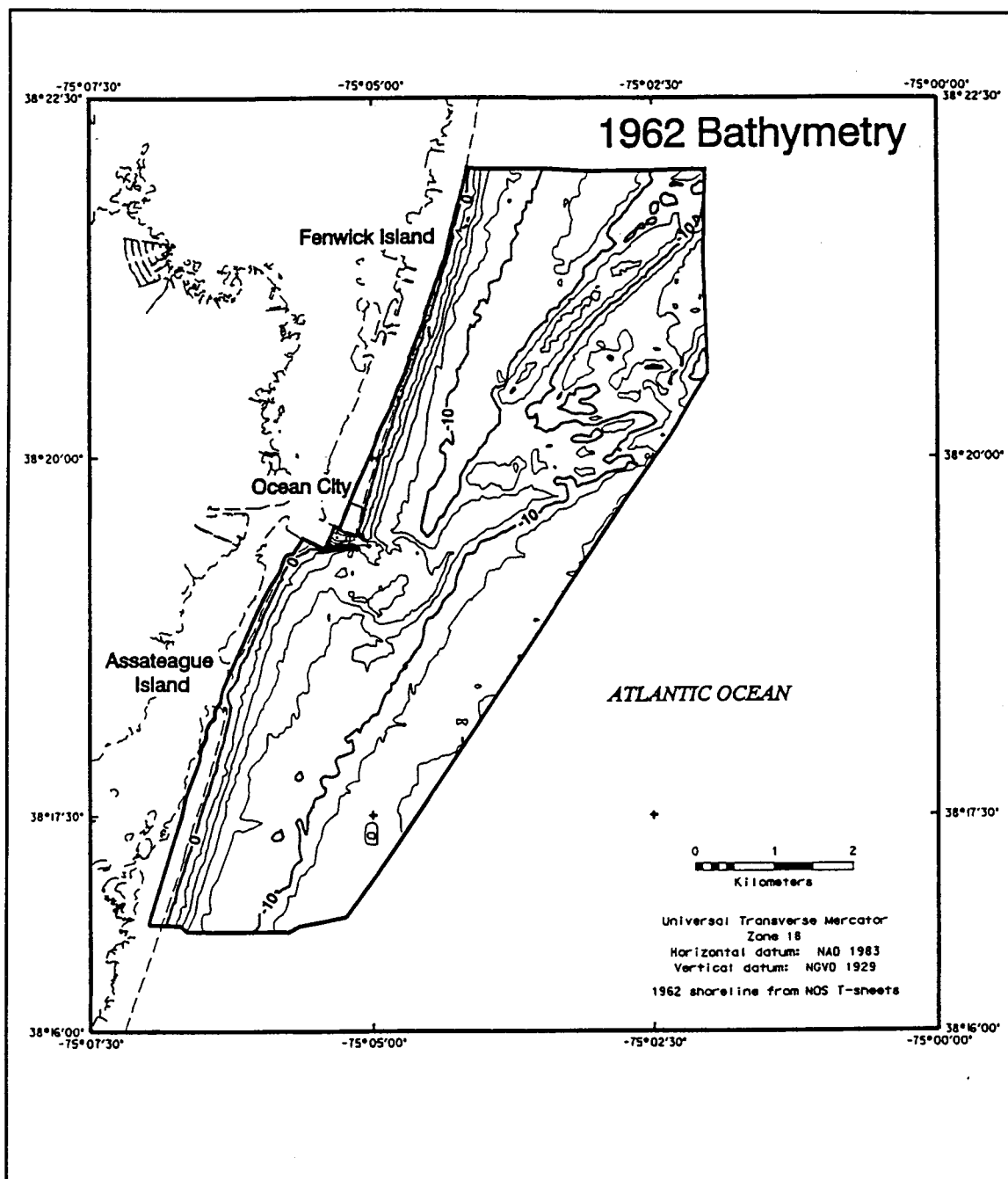


Figure 59. 1962 ebb shoal morphology. Note shoal configuration becoming crescentic and slightly offset to the south.

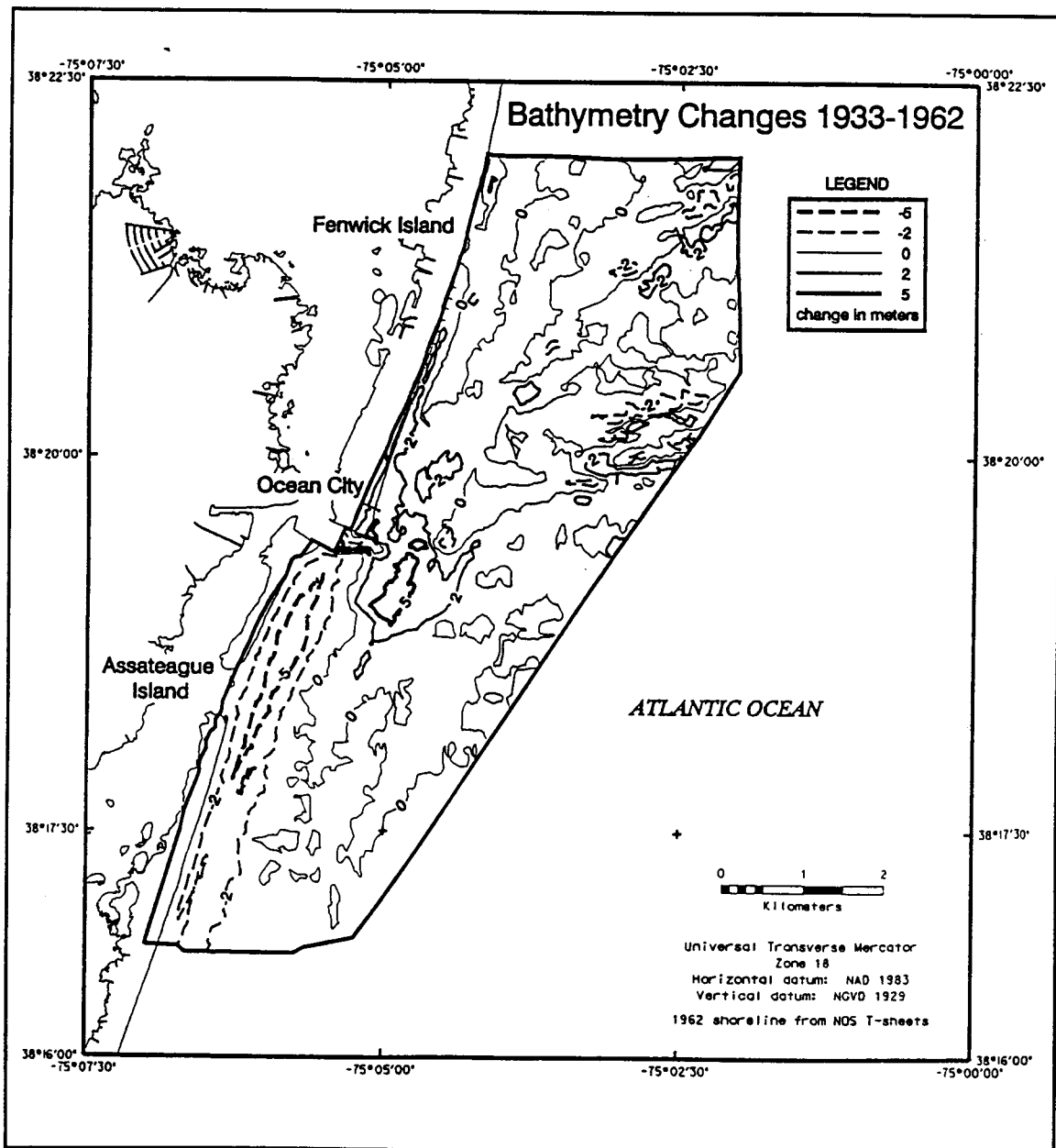


Figure 60. Bathymetric difference map from 1933-1962. Note vertical accretion of approximately 5 m along the interior, and 2 m along the southern flanks.

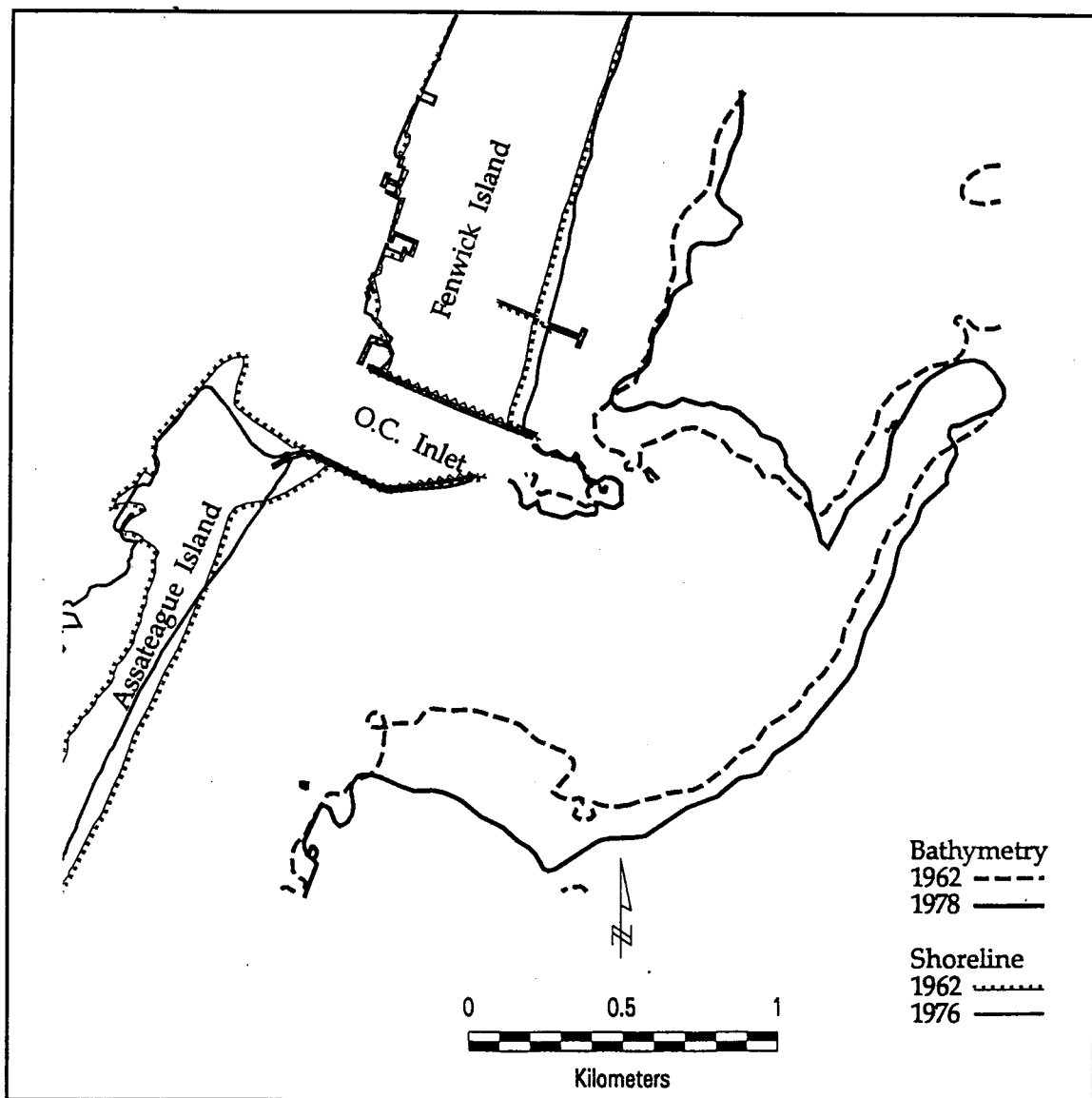


Figure 61. -7 m contour changes from 1962-1978. Note increased width along the southern boundary.

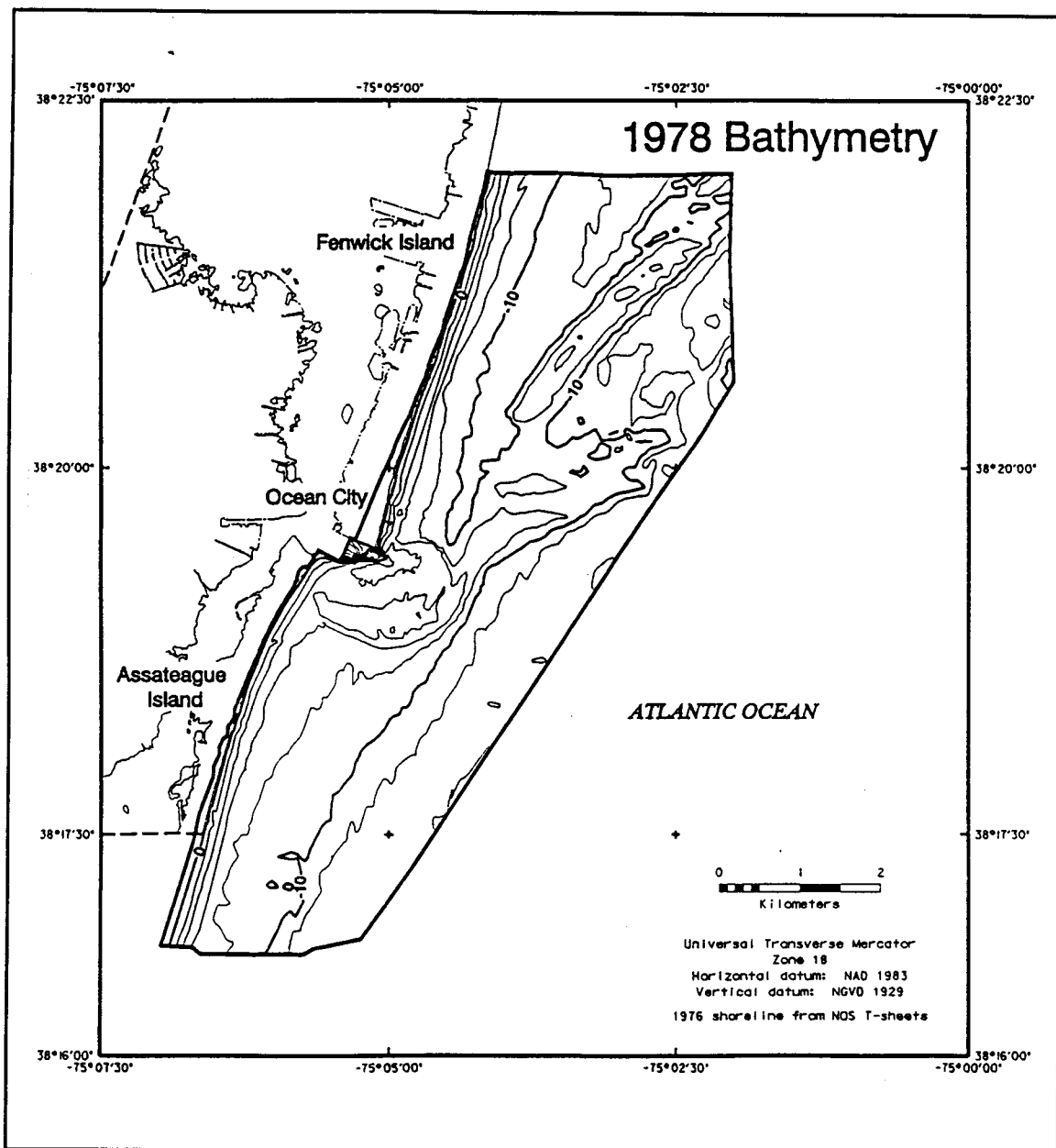


Figure 62. 1978 bathymetry showing a more pronounced extension of the ebb shoal moving toward attachment to the downdrift shoreline of northern Assateague Island.

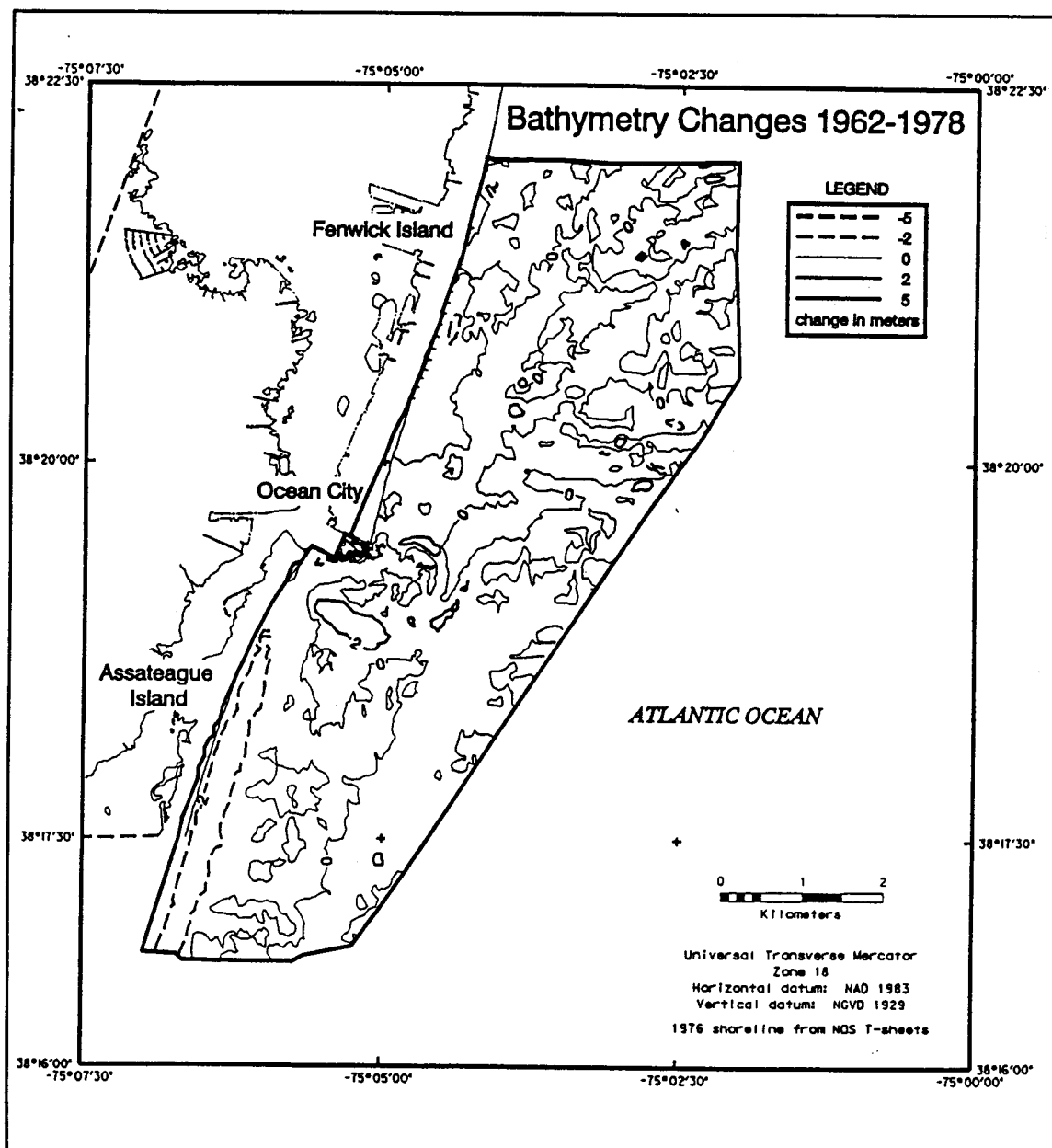


Figure 63. Bathymetric difference map from 1962-1978. Note changes occurring predominately along the outer portions of the shoal.

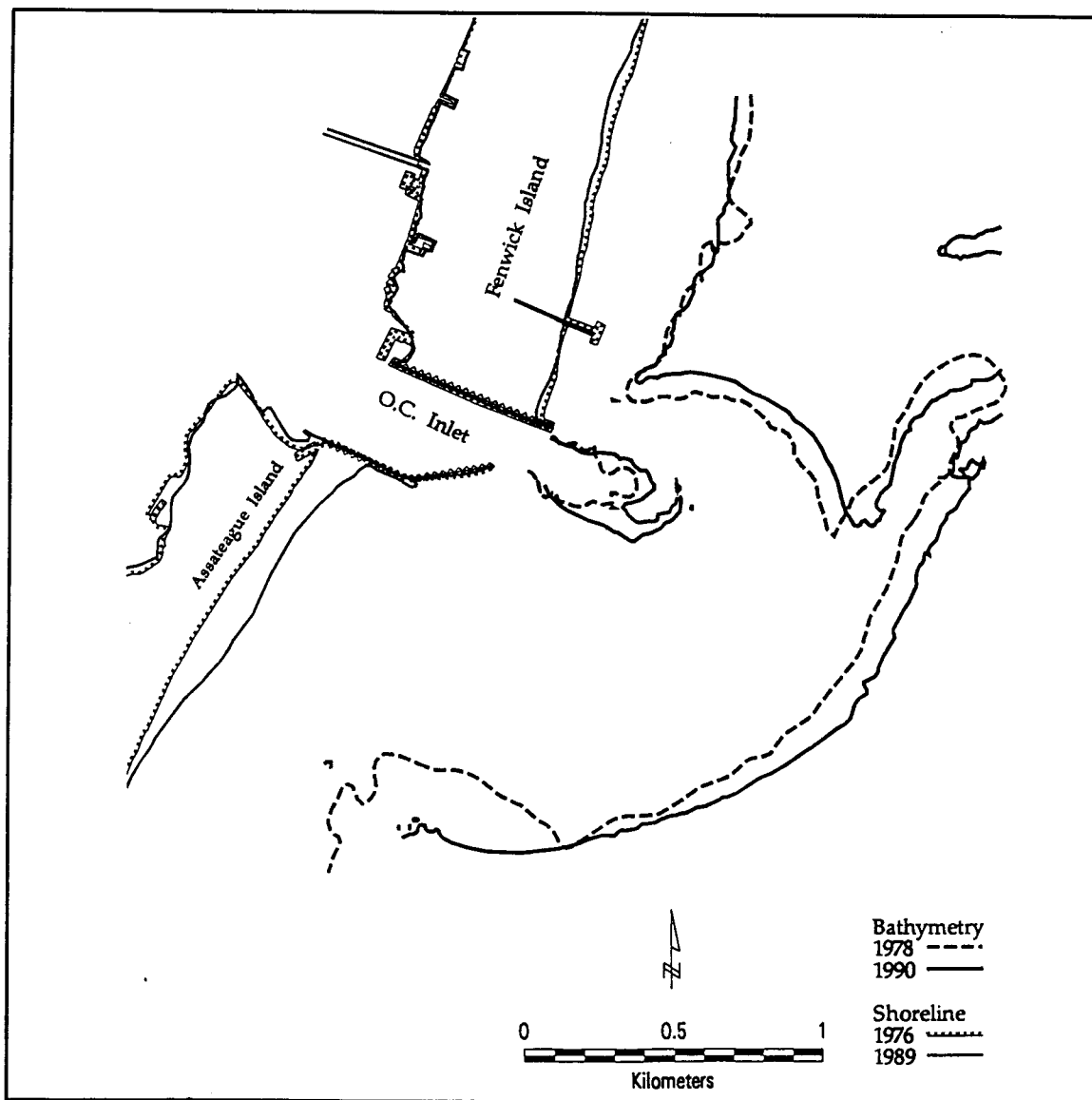


Figure 64. -7 m contour changes from 1978-1990.

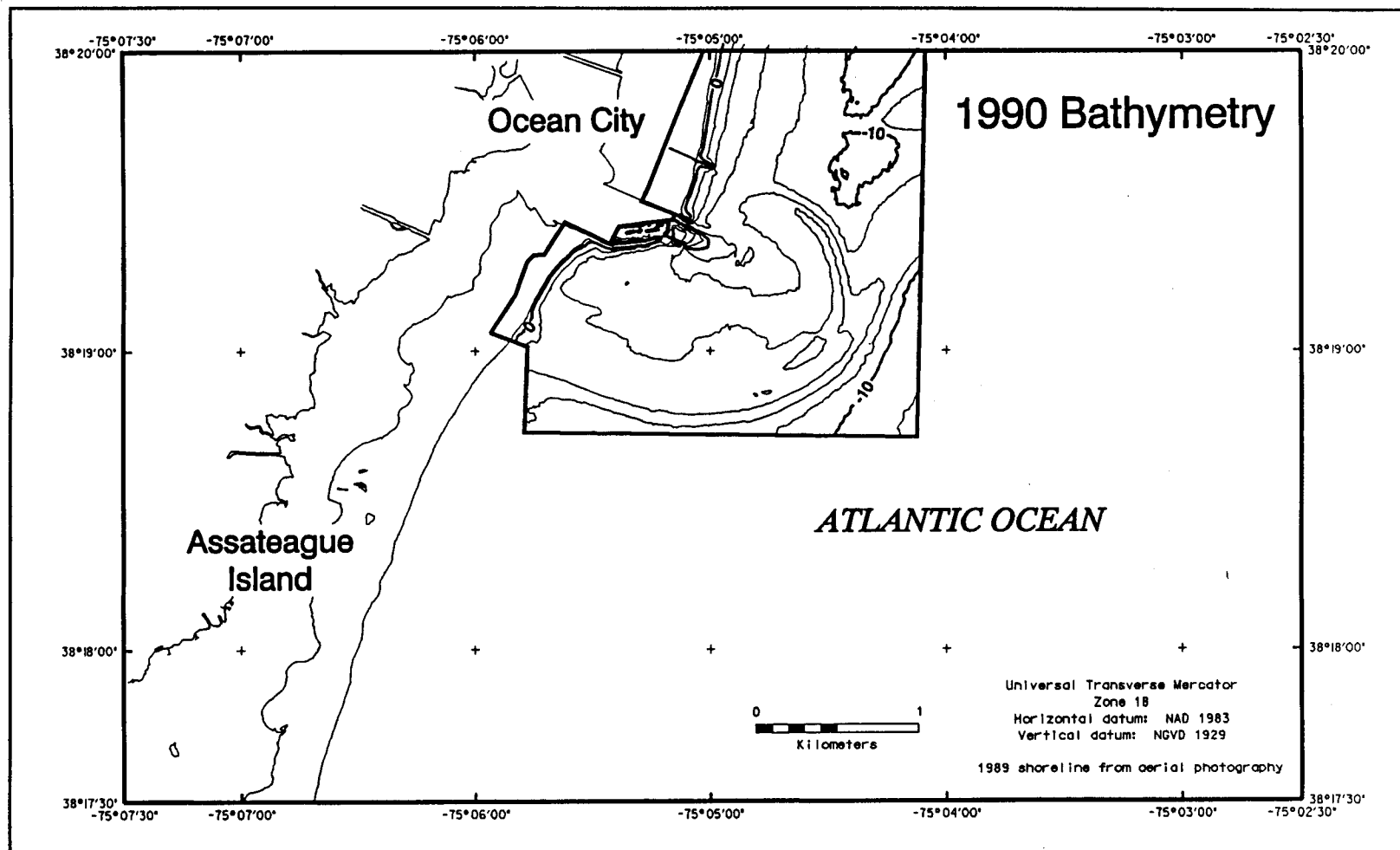


Figure 65. 1990 bathymetry showing attachment to northern Assateague Island.

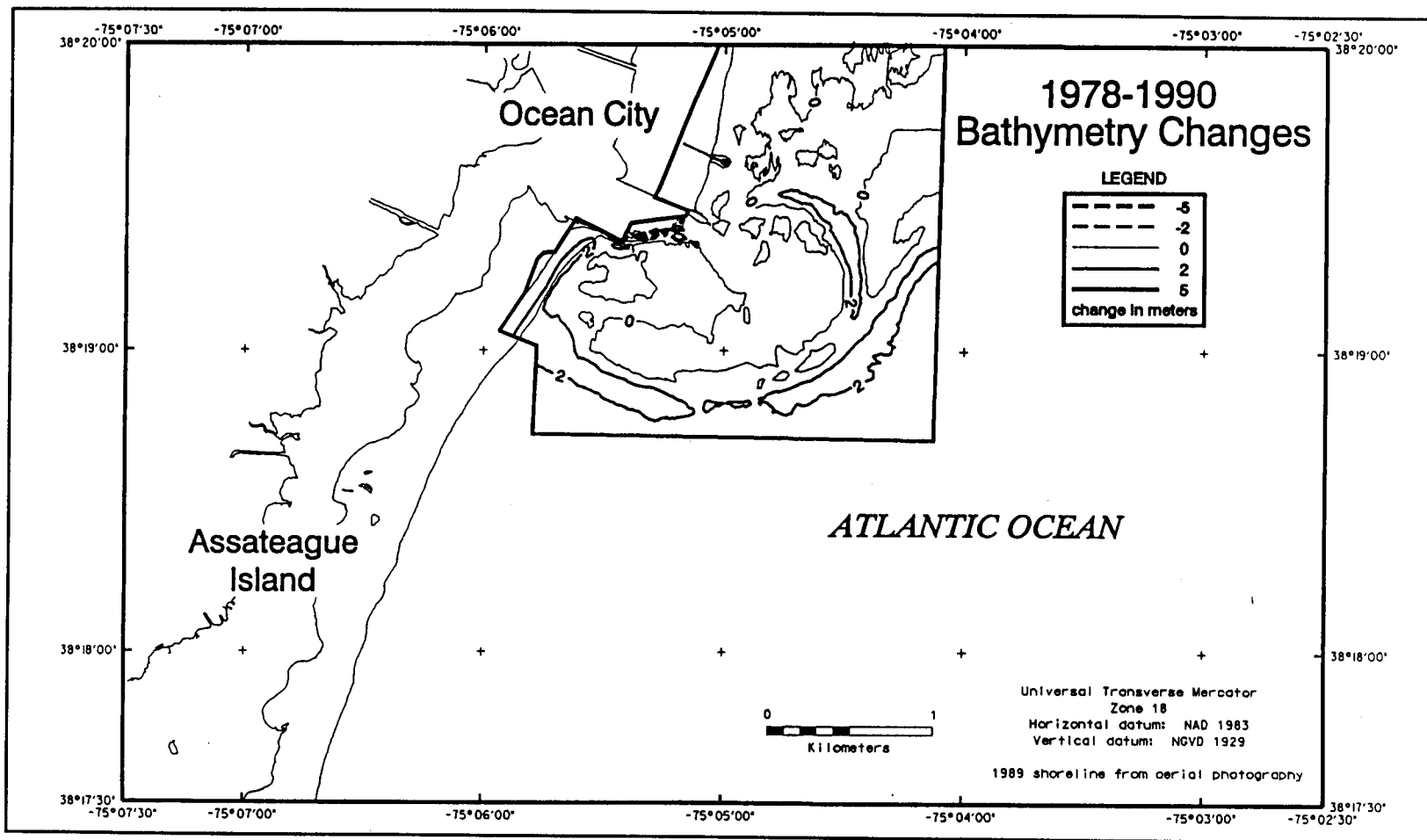


Figure 66. Bathymetric difference map from 1978-1990. Note 2 m vertical accretion along the outer lobe, and linear shoal complex, and attachment to the downdrift shoreline.

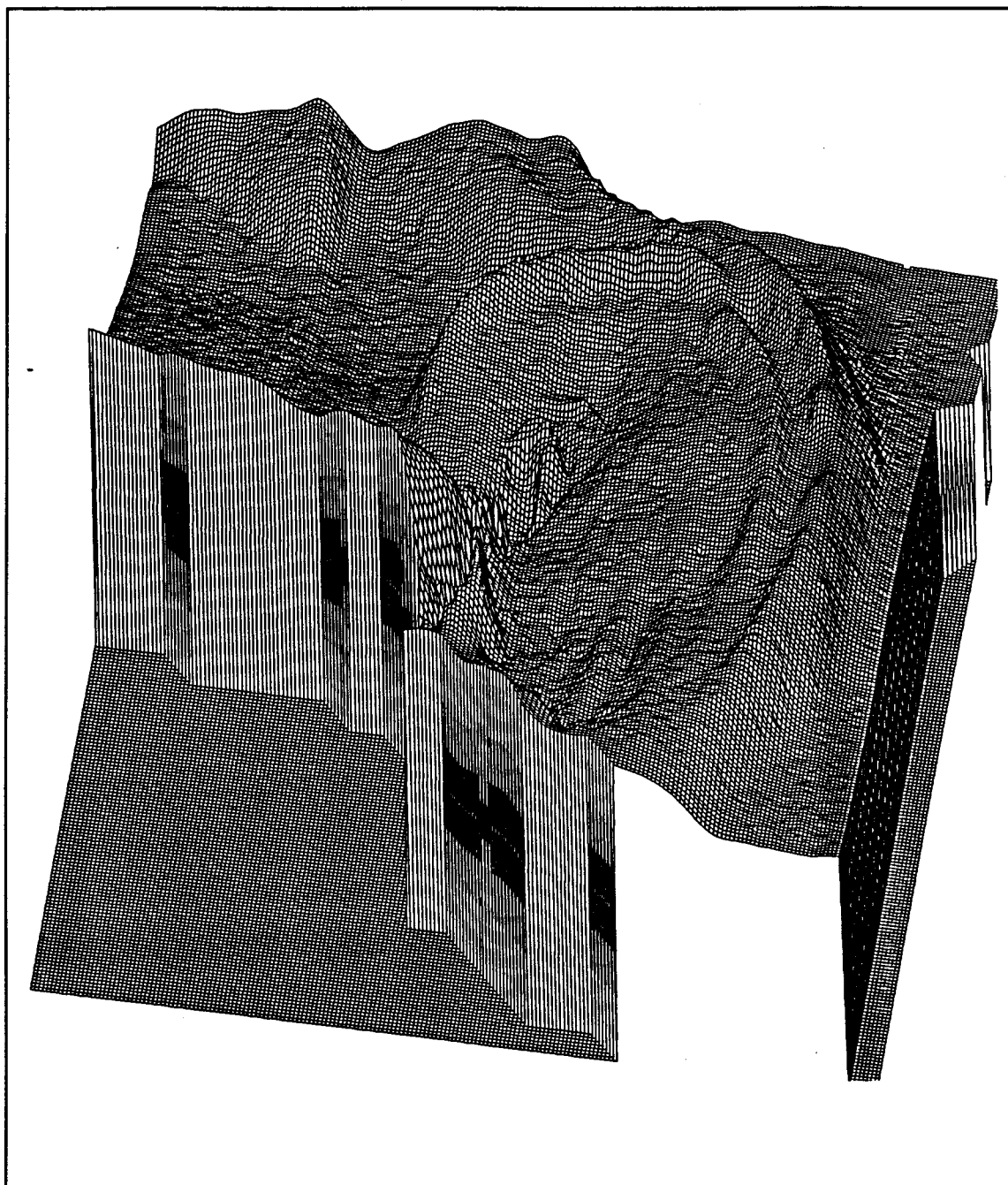


Figure 67. 1990 three dimensional view from the south-southwest. Notice the double bar along the southeastern edge joining near the shoreline.

The rate of shoal volume change between 1933-1962, 1962-1978, and from 1978-1990 was approximately 151,000 m³/yr, 101,000 m³/yr, and 150,000 m³/yr respectively. Comparisons of ebb shoal lateral development for the first 30 years (1933-1962), and the last 30 years (1962-1990), shows a decrease in offshore growth from approximately 1140 m to 170 m, and lateral growth from 1330 m to 570 m. Accretional vertical changes as much as 5 m occurred from 1933 to 1962. Since 1962, maximum vertical growth was 2 m, and this was restricted to the outer edges of the shoal.

7 Discussion

Shoreline Change

The formation of Ocean City Inlet along this particular segment of northern Assateague Island, can be linked to a number of geomorphic and meteorological factors. The northern portion can be described as a low-lying overwash area. These segments of barrier islands are, by definition, frequently overwashed, have little relief, with dune elevations less than 1.5 m, and are covered with meadow or grassland vegetation adapted to frequent sand burial and flooding by overwash (Leatherman 1988). Another indication of the low relief nature of this part of Assateague Island is given by Truitt (1968). He used the term "inlet shallows" to describe the northern 1.2 km. Again, this descriptive term was used to characterize this area of frequent overwash and ephemeral inlet locations. Consequently, this low lying portion of Assateague Island was an ideal location for new inlet formation.

Examination of historical engineering records in the 1920's reveals that numerous groins were constructed along the southern portion of Fenwick Island. Before groin construction in the 1920's, the average shoreline retreat rates were -2.70 m/yr (1850-1908), increasing to approximately -4.23 m/yr (1908-1933) near the origin of the inlet breach. North of this location shoreline retreat rates were -1.2 m. Quantifying what impact these structures had on increasing shoreline retreat rates is not possible due to the lack of available information. However, these structures were undoubtedly trapping sediment along their northern portion, and reducing available sediment downdrift.

Storm events, both tropical and extratropical, heavily impact this area during the winter months, and were shown to be the dominant mechanism of sediment transport (Douglas 1985). The 1933 hurricane passed just south of this area, with its most powerful and damaging sector (northeastern quadrant) passing over Assateague Island. Elevated storm surge levels caused by this storm would have inundated the island, in a scenario similar to that which created Sinepuxent Inlet in 1920. This newly created inlet became a sink for sediments from the adjacent shorelines of Fenwick and Assateague Island.

Due to pressure from the commercial finfish and shellfish industry and other business groups, the USACE decided to stabilize the inlet. Stabilization was initiated primarily to improve navigational conditions and prevent natural closure from longshore sediment transport. After north and south jetty construction was completed in 1935, the adjacent shorelines of southern Fenwick and northern Assateague began to respond. Initially, the southern shoreline of Fenwick Island began building seaward through sediment becoming trapped along the adjacent north jetty. From 1935-1937, 114,000 m³/yr of sediment were being trapped within the north fillet of the jetty (Dean et al 1979). By 1947, the north jetty was impounded to capacity. Dean and Perlin (1977) state; "an average of 76 percent of the net material arriving at the north jetty would be bypassed".

Sediment arriving at the inlet was becoming trapped either within the landward flood tidal delta or the inlet side adjacent to the north jetty, and/or being jetted offshore by the strong ebb currents to become deposited within the ebb tidal delta.

The post-inlet average shoreline retreat rate for the northern 6.5 km of Assateague, increased from -6.37 m/yr (1933-1942) to -9.41 m/yr (1942-1962), and lessened to -4.42 m/yr in 1976. This reduced rate from 1962-1976 was influenced by USACE dredging activities during post recovery efforts from the 1962 Ash Wednesday storm. However, the average rate of shoreline retreat for the entire period (1933-1976) was -9.16 m/yr. Since 1976, shoreline changes as compared to the 1933 shoreline have lowered to the present rate of -6.30 m/yr.

Ebb Tidal Delta Growth

Following formation of Ocean City Inlet in August 1933, an extensive ebb shoal began forming seaward of the inlet. Initially strong tidal flows (both ebb and flood) associated with the new inlet opening, and existing large tidal prism, moved large quantities of sediment offshore to become deposited into the ebb shoal. The ebb shoal platform grew from sediment within the inlet throat, derived from scouring of the inlet entrance, and alongshore suspended sediment movement during hurricanes and other large extratropical storm events. These storm events have also had a great deal of influence on the ebb shoal's morphology. As these storm waves break along the outer edge of the terminal lobe, they tend to place the bottom material into suspension, push the bar back towards and/or along the shore, and pile some of the material back on the ebb shoal platform. At the same time, some of this material is pushed sideways, flattening and elongating the face of the shoal where the predominant wave activity is focused. For example, the prevalent northeasters, strike the ebb shoal on its northeastern point, flattening, pushing, and elongating the shoal in a southerly or southwesterly direction, thus moving this material toward northern Assateague Island.

Between 1933-1962, growth in the offshore (eastern) direction continued to a point at which incident wave energy became greater than that of tidal energy (approximately 1100 m). At the same time, northeaster storms and other high energy events were contributing to a southern downdrift offset, and more streamlined ebb shoal platform. Ebb shoal volume increased steadily from 1933-1962, slowed from 1962-1978, and again rose steadily from 1978-1990. The slower shoal growth from 1962-1978 may mark the beginning of sediment bypassing. This time period had the greatest increase in longshore growth across the ebb shoal. This lateral movement of sediment was probably the initial stage before shoal attachment to Assateague Island. After 1978, the rate of shoal volume change per year increased to approximately 150,000 m³/yr, equalling the pre 1962 rate. This additional increase in shoal volume may be due to beach nourishment operations undertaken in 1988. Ebb shoal accumulation of approximately 8.0 million m³ has occurred over the last 60 years, for an average rate of 138,298 m³/yr.

The perpetual growth, in conjunction with storm events, has led to an attachment or welding of the ebb shoal to northern Assateague Island. Bathymetric and shoreline data presented in this report, along with historical oblique aerial photography suggest this attachment occurred after 1976, probably around 1980. The 1980 date is based on visual interpretation from aerial photography of a distinctive shoreline bulge, where the shoal has welded itself to the downdrift shoreline. This physical attachment will allow this morphologic feature to act as "*a sand bridge*", bringing much needed sediment to northern Assateague. The sediment bypassing mechanism may have involved a two stage process; initially through ebb flows pushing sediment offshore, and then bar-bypassing processes of formation, and landward migration of swash bar complexes, via wave induced sediment transport across and along the shoal platform to northern Assateague.

In figure 68, comparisons of the average shoreline change rate and ebb shoal volume growth from 1850-1990 show a reduced shoreline change rate after 1976, corresponding to reduced volume growth of the ebb shoal. This period of sediment bypassing has resulted in accretion along the northern portion of Assateague Island since 1976, as evidenced in figure 69.

The evolution and growth of the ebb shoal has been uninterrupted by any human activity since jetty construction. The jetties have allowed the ebb shoal to develop farther seaward, due to entrainment of tidal flow during the ebb tidal cycle. This has undoubtedly provided greater tidal flows through the inlet, helping flush any accumulation of excess sediment within the inlet throat both in a landward and seaward direction. Grosskopf and Resio (1987), conducted a wave refraction study on the effects of dredging activities across the ebb shoal. Specifically, their study examined the effect of partial or total removal of the ebb shoal and the impact this dredging activity would have on the adjacent shorelines of Fenwick and Assateague Island. Their results found the ebb shoal to be completely unsuitable for mining. Removal of even a small amount of sand from this shoal, could potentially result in serious shoreline erosion of northern Assateague. The predominate dredging activities have all occurred in the navigation channel near the junction of Sinepuxent and Isle of Wight Bays (Commercial Harbor area). Over the last 60 years, this natural evolution of the ebb shoal (uninterrupted growth due to the absence of dredging on the ebb shoal proper) has allowed it to evolve into its present configuration.

Left undisturbed, the future evolution and growth of the shoal will be predominately in a lateral or downdrift direction from the inlet. This shoal will continue to shield northern Assateague from maximum storm wave energies, and at the same time re-establish and provide for longshore littoral drift movement downdrift from the inlet, and towards the inlet (through wave refraction). The amount of sediment bypassed to northern Assateague Island will vary based on yearly storm conditions, dredging activities, linear shoal contributions, and future beach re-plenishment operations along Ocean City's shoreline. However, an attempt at determining average annual sediment bypassing rates is given in Table 8 and figure 70, based on the following assumptions:

1) Annual net southerly transport rate of 212,800 m³/yr (Douglas 1985). Douglas's (1985) transport rate was deemed the most appropriate for the following reasons: The ebb shoal has grown at an annual rate of approximately 138,298 m³/yr, which is approximately equal to Dean

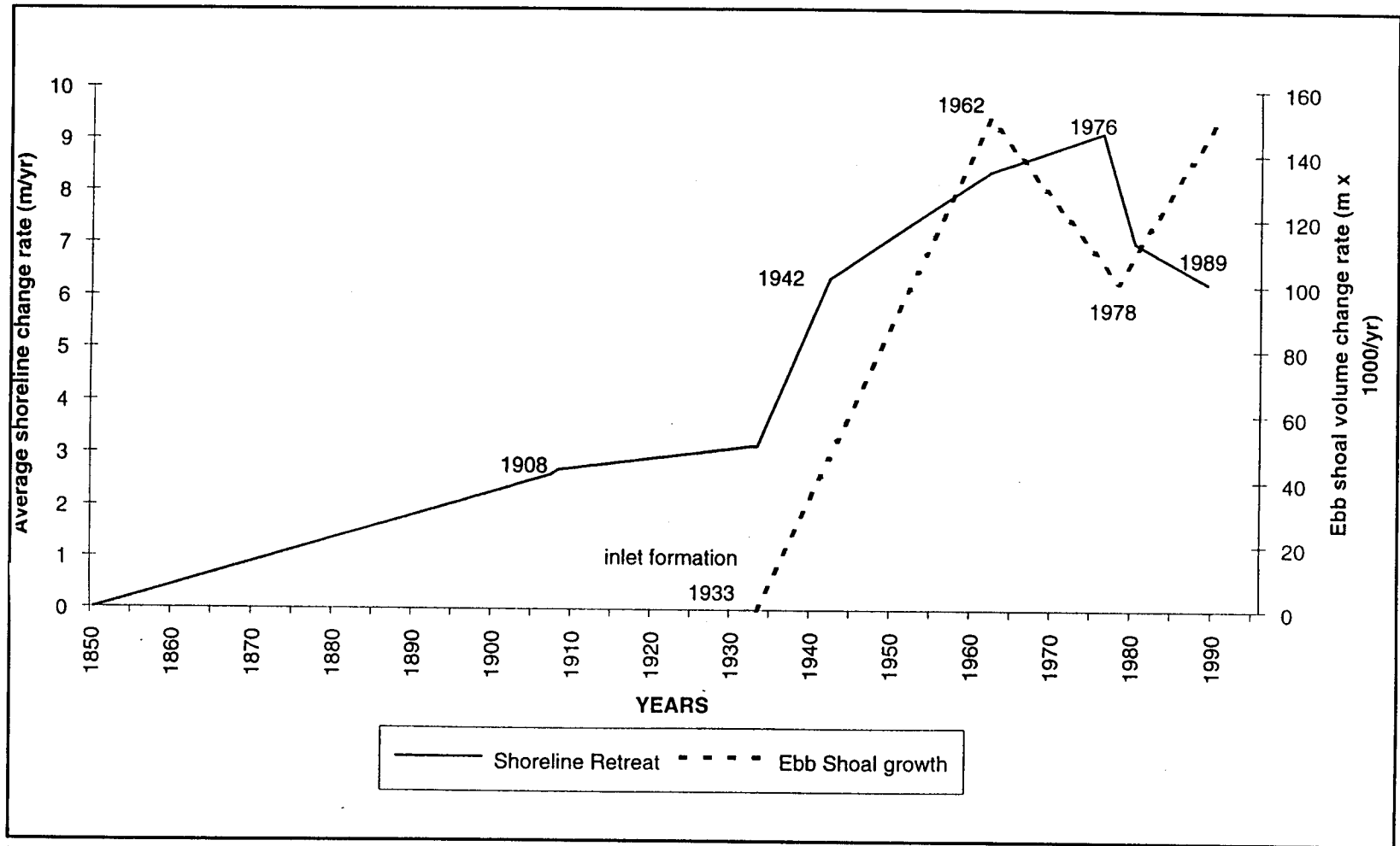


Figure 68. Comparison of average shoreline retreat and ebb shoal volume growth from 1850-1989.

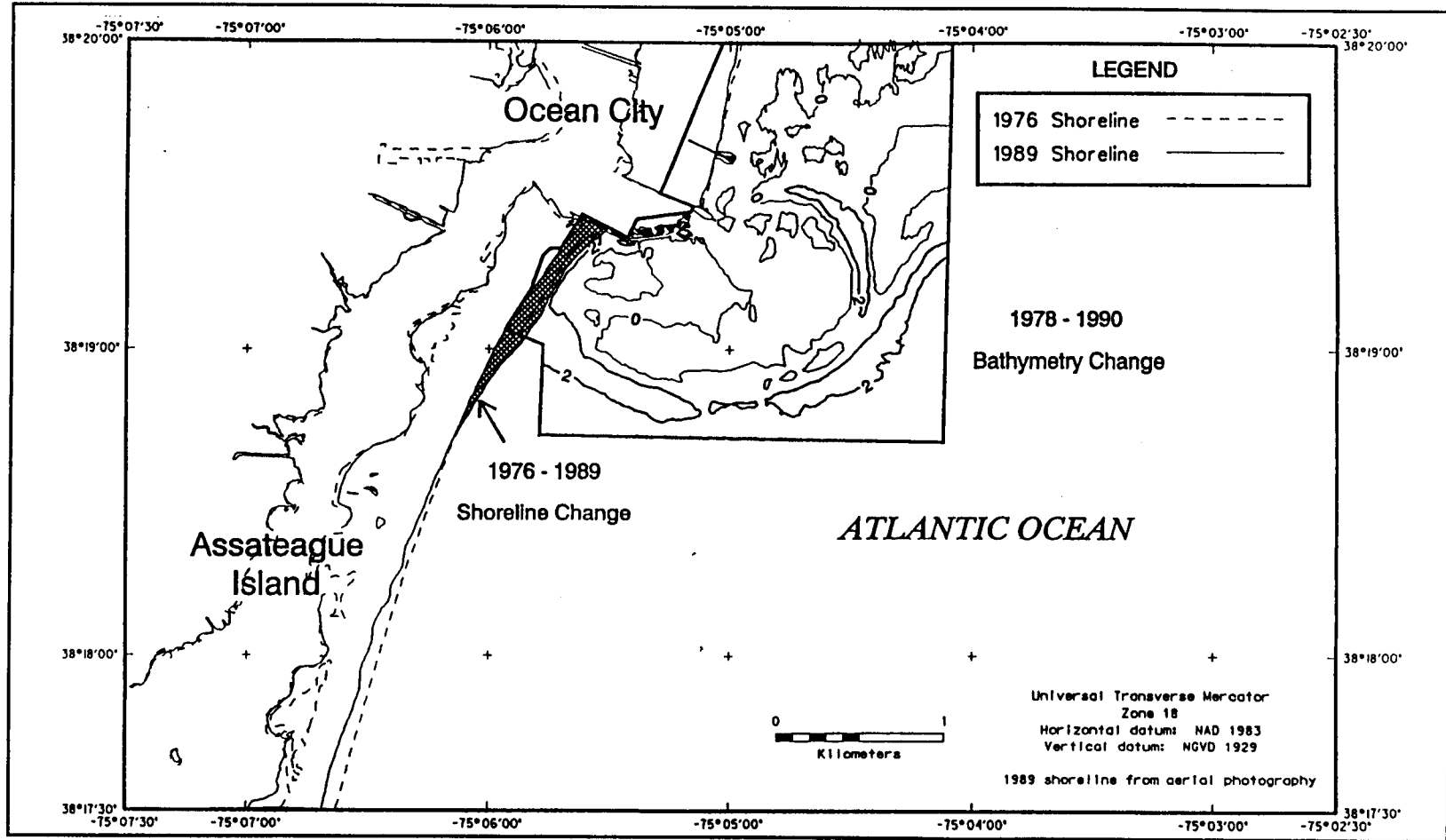


Figure 69. Shoreline changes from 1976-1989, and 1978-1990 bathymetry changes, note the shoreline advancement in conjunction to welding of the ebb shoal to the shoreline.

Table 8. Estimate of sediment bypassing quantity to northern Assateague Island. Reference figure 70 for locations of letters A – E			
Parameter	Rate (m/yr)	Percent of Net	model inputs
Net littoral drift	212,800	n/a	+
Ebb shoal	138,298	65	+
Flood shoal	18,536	9	+
1.6 km stretch of	87,000	41	+
Littoral drift reversal	39,900	19	-
¹ Other	8,512	4	+
¹ refers to inlet throat, or north jetty	TOTAL BYPASSED = approximately 25 percent of NET approximately 50,000 m ³ /yr		

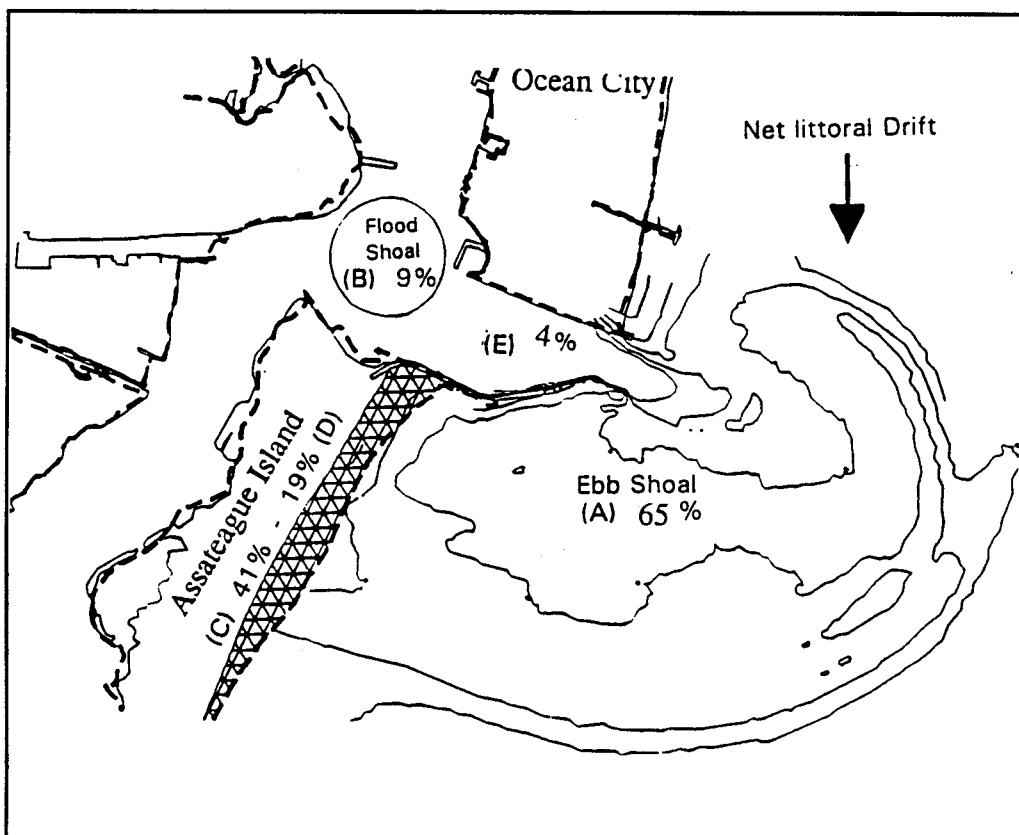


Figure 70. Sediment bypassing parameters from Table 8, depicting their location along the study site.

et al's (1979) estimates (114,000 m³/yr - 152,000 m³/yr). The flood tidal shoal has continued to grow over the same time period capturing a portion of the net littoral drift. Therefore, the net rate of 212,800 m³/yr was chosen to be the most representative of the total amount of sediment arriving at the inlet.

2) Average annual ebb shoal growth of 138,298 m³/yr. Based on total ebb shoal volume of 7,883,000 m³ from 1933-1990 (57 years).

3) Average annual flood tidal shoal growth of 18,536 m³/yr. Based on a total flood shoal volume of 760,000 m³ from 1931-1972 (41 years) (Dean et al 1979). This data was one component used for a sediment budget analysis at Ocean City Inlet.

4) Average annual sediment transport through and over the south jetty of 39,900 m³/yr. Based on the average between 34,200 m³/yr and 45,600 m³/yr under normal wave conditions. This data was used in a report by Bass et al. (1994), cited from Dean and Perlin (1977). Since the south jetty rehabilitation project was complete in 1985, this rate represents the sediment quantity normally transported towards the inlet by wave refraction around the ebb shoal (i.e. no sediment bypassing).

5) Average annual accretion rate for northern Assateague Island of 87,000 m³/yr. Based on the progradation of the northern 1.6 km shoreline adjacent to the south jetty from 1976-1989. Assuming that .09 m² of beach change is equivalent to about .76 m³ of sand, results in an overall estimated accretion of approximately 1.13 million m³ of sand (March 1976-April 1989).

The shoreline adjacent to the inlet (approximately 4 km) should continue to prograde seaward due to natural sediment bypassing and the sand tightening of the south jetty in 1985. This rate of progradation should continue to lower the shoreline retreat rate by approximately 2-3 m per 10 years. This reduction in shoreline change rate is based on the trend of a 1 m decrease from 1980-1989 in relation to the 1933 shoreline; sediment contributions from the linear shoal complex (overtopped during seaward progradation of the ebb shoal); the subaqueous bar welded to the downdrift shoreline of northern Assateague Island; past (1988-1991) and future beach nourishment activities that are planned for a four year interval.

8 Conclusions and Recommendations

- 1) Engineering activities have had both negative and positive impacts on northern Assateague Island. Negative impacts include inlet stabilization through construction of jetties in 1935. Average shoreline retreat rates escalated from -3.2 m/yr (before inlet stabilization) to a maximum of -9.0 m/yr in 1976. Positive impacts include sand tightening of the south jetty in 1985, and beach replenishment activities in 1988.
- 2) Cumulative shoreline change rates for northern Assateague Island (1933-1989) shows reduced retreat rates adjacent to the inlet since 1976, from a high of -9.0 m/yr to the present -6.5 m/yr (inlet south 6 km). Present erosion rates of -2 m/yr were measured for the remaining 13 km.
- 3) Ebb shoal volume increased steadily from 1933-1962 (151,000 m³/yr), slowed from 1962-1978 (101,000 m³/yr), and again rose steadily from 1978-1990 (150,000 m³/yr). Ebb shoal volume since 1933 has averaged approximately 140,000 m³/yr.
- 4) The ebb shoal has become physically attached to northern Assateague Island approximately 600m from the south jetty. Since 1976, the northern 1.6 km shoreline of Assateague Island has shown an accretional trend, accumulating approximately 1.13 million cu m of sand, averaging approximately 87,000 cu m/yr.
- 5) Natural sediment bypassing is presently occurring at an estimated rate of approximately 50,000 m³/yr, which is approximately 25 percent of the total net littoral drift.

The National Park Service (NPS) is responsible for long term protection and preservation of Assateague Island National Seashore. Currently, the NPS is faced with a critical policy dilemma. The problem focuses on whether to interfere with present shoreline dynamics, or face losing a portion of northern Assateague Island. The northern end is habitat for a variety of rare species, including the federally listed threatened piping plover. Preservation policies of the NPS usually emphasize the maintenance of *natural* processes. However, human intervention may be justified because of the following reasons: 1) much of the observed impact is due to human activities (i.e. inlet stabilization); 2) the stewardship role of the NPS in conserving nationally significant natural resources; 3) the requirement that all federal agencies activities promote the recovery of threatened and endangered species and associated critical habitat; and 4) the need to protect submerged and terrestrial resources landward of the barrier island.

Understanding the local sedimentological and geomorphological processes will provide improved planning and design criteria for physical monitoring programs, and establish the importance of the ebb shoal in providing erosion control on adjacent northern Assateague Island. To maintain northern Assateague's present level of protection, and provide for increased preservation of the northern most section of Assateague Island, the following recommendations are

made: (1) Any dredging activities planned for the ebb shoal should be discouraged. This would slow or completely halt natural sediment bypassing to northern Assateague Island; (2) Any sediment removed by dredging activities within the inlet throat section or back bay areas, should be placed downdrift of the inlet. This additional sediment could be placed within the oceanside intertidal area to be redistributed naturally by wave action. Hastening shoreline retreat along the remaining 6.5 km of shoreline. Designed properly, the beachfill could be configured to restore critical island width, with minimal disturbance to the natural habitat; (3) Design of a beachfill project that would widen the northern end, and allow natural re-establishment of low-lying dunes, providing the *jump start* that is desperately needed after 60 years of sediment starvation.

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