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Morphodynamics of Tidal Inlet Systems in Maine

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ABSTRACT

The occurrence of tidal inlets along the coast of Maine is tied closely to the structural geology and glacial history of this region. Most of the inlets are found along the southern arcuate-embayment shoreline where sand sources, consisting of glaciomarine sediments and other glacial deposits, were sufficient to build swash-aligned barriers between pronounced bedrock headlands. Along the peninsula coast of Maine, tidal inlets also occur at the mouths of the Kennebec and Sheepscot Rivers where large quantities of glaciofluvial sands were deposited during deglaciation. The remainder of the southeastward facing coast was stripped of its preglacial sediment cover by the southerly moving glaciers. The thin tills that were left behind yield little sand and, thus, barriers and inlets are generally absent.

Small to large-sized inlets (width = 50-200 m) in Maine are anchored next to bedrock outcrops and are bordered on their opposite sides by sandy spits. Despite the ubiquitous name "river inlet," they normally have little fresh water discharge compared to their salt water tidal prisms. The backbarriers of these inlets are expansive and would produce relatively large tidal prisms if high *Spartina* marshes had not filled most of the region, leaving little open water area. While all the inlets have well developed flood-tidal deltas, only the large inlets have well formed ebb-tidal deltas. At small and medium-sized inlets (width <100 m), ebb deltas are absent or poorly developed.

The hydraulics of Maine's inlets are controlled by their size. Small and medium-sized inlets are dominated by flood tidal currents due to a longer ebb than flood duration. This duration asymmetry is the result of both frictional effects and a truncation of the tidal wave by the shallowness of the entrance channel. These conditions are much less apparent at large inlets which have stronger ebb tidal currents. The ebb dominance of these inlets can be explained in terms of inlet efficiency and filling characteristics of the backbarrier. The three estuaries discussed in this paper have very different freshwater discharges and tidal prisms. The mouths of the Kennebec and Sheepscot Rivers are normally tidally dominated (excepting spring freshets), while the Saco River mouth exhibits stratified flow and more estuarine conditions.

Sediment dispersal at Maine's inlets is characterized by discrete, well-defined sand recirculation cells. These systems can include just the inlet and adjacent spit platform or it can involve the ebb-tidal delta and a 2-3 km length of adjacent beach. At small and medium-sized inlets, sand is also transported into the backbarrier by the dominant

flood tidal currents. This pattern of sand transport at inlets in Maine differs markedly from inlets along open ocean, barrier island coasts (East and Gulf Coasts) where most of the sand eventually bypasses an inlet to the next downdrift barrier.

INTRODUCTION

Tidal inlets in Maine owe their presence to swash-aligned barrier systems formed by Holocene sedimentary processes. They are concentrated between bedrock headlands along the arcuate embayment coast from Ogunquit to Cape Elizabeth because of abundant sand supply in this part of Maine (Fig. 1). North and south of this region there are numerous harbors and

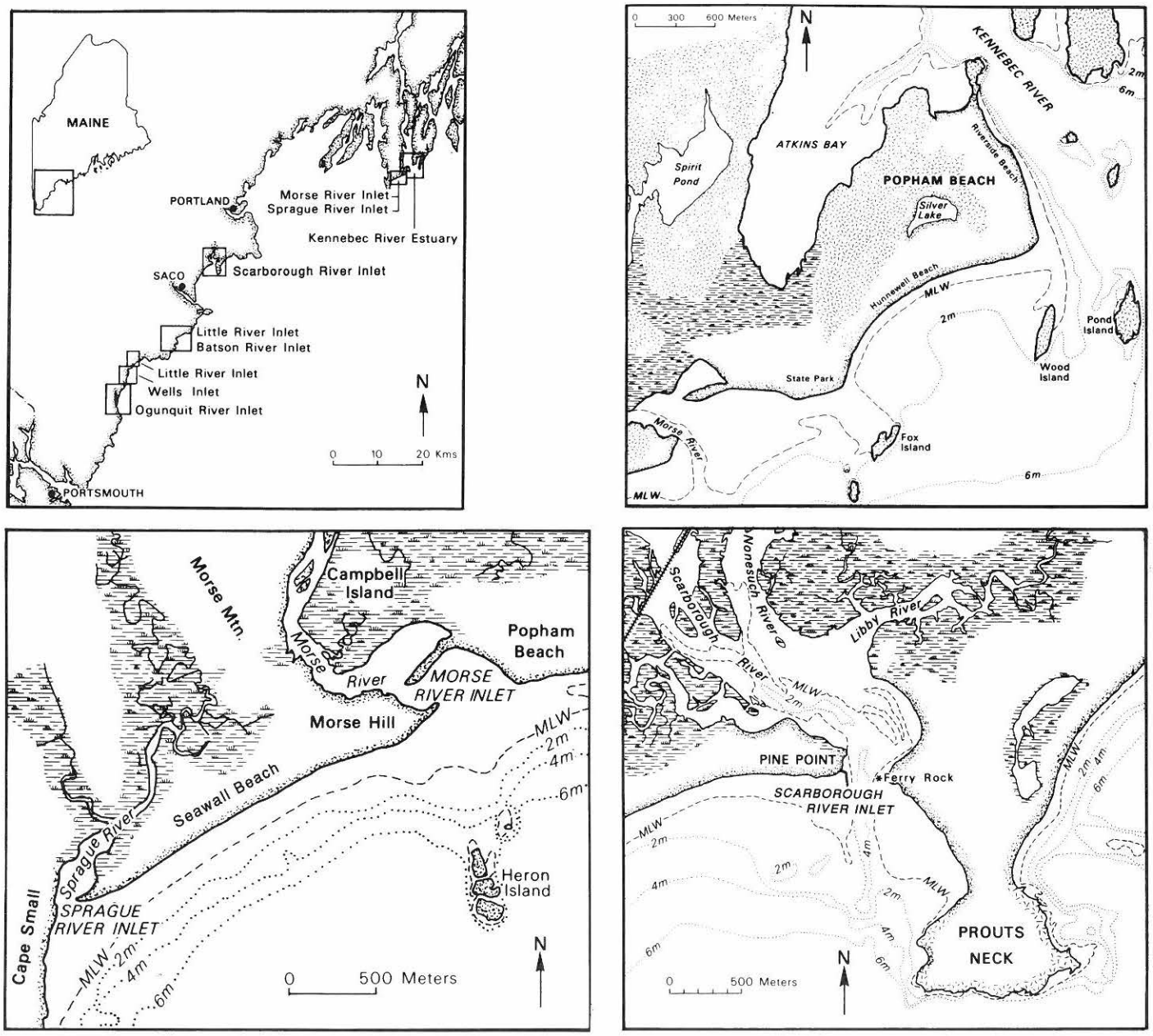

Figure 1. Location of field studies.

bays, but little sand accumulates along the coast. Two major exceptions to this pattern are the tidal inlets found along the sandy mouths of the Sheepscot and Kennebec River Estuaries east of Casco Bay.

Due to the effects of glaciation and the coastal bedrock headlands, Maine's inlets differ markedly from those that occur along straight barrier shorelines such as those found on the coasts of Texas, East Friesian Islands, West Germany, and Egypt (Nile Delta). Drift-aligned shorelines, including those of the U.S. East Coast (south of New England) and Gulf Coast, generally exhibit a dominant longshore transport direction; thus inlets and their ebb-tidal deltas along these coasts generally experience channel migrations and processes of sediment bypassing. In contrast, the inlets along Maine's moderately to highly indented coast are bedrock-controlled, have relatively stable inlet throat positions, and exhibit discrete well-defined sand recirculation patterns.

Although the beaches of Maine have been studied in considerable detail, including their sedimentology, morphology, and shoreline changes (Fink et al., in press; Nelson, 1979; Trudeau, 1979), botany (Trudeau et al., 1977; Nelson and Fink, 1978), and

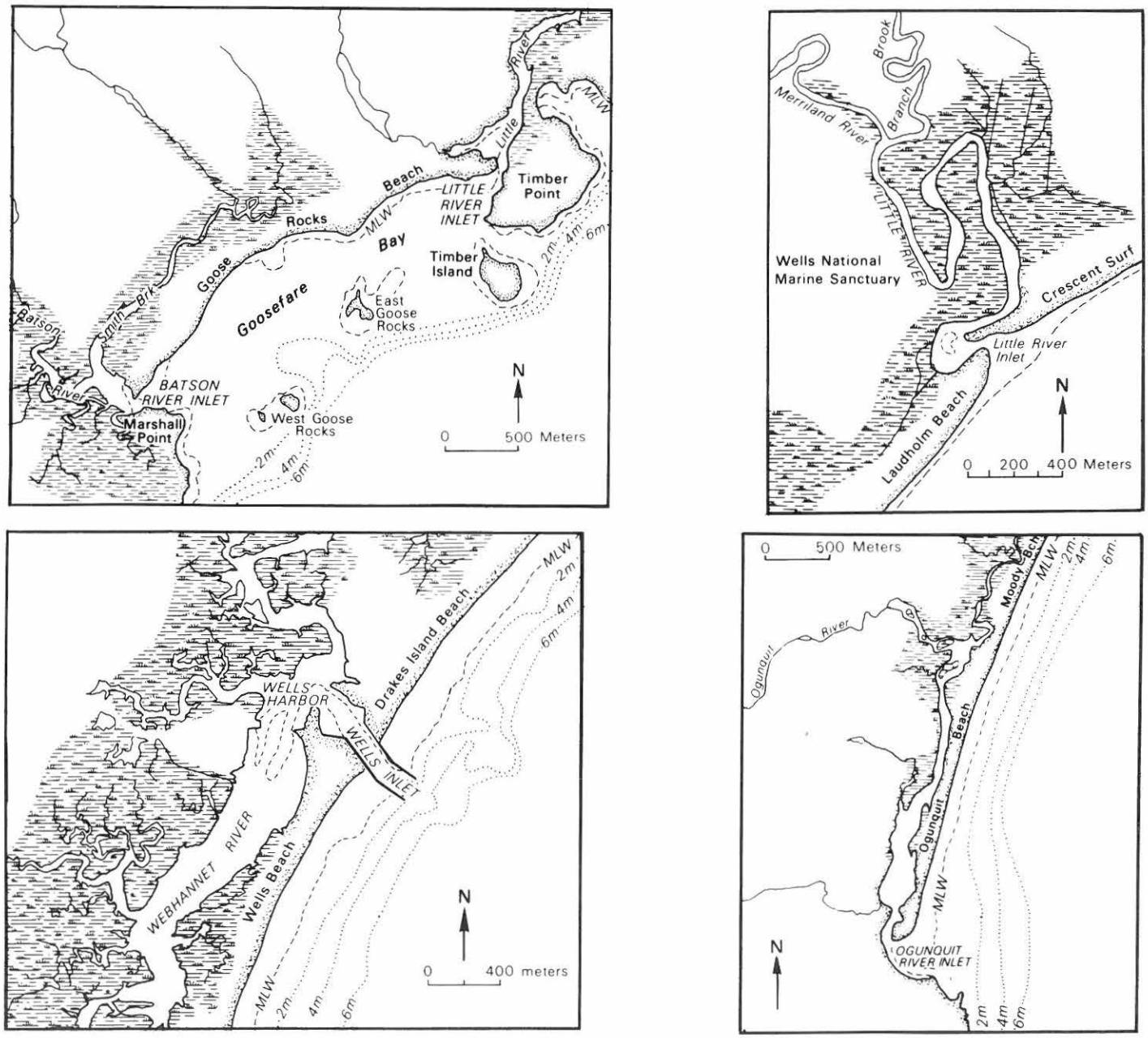

Figure 1 (Continued). Location of field studies.

dune and barrier dynamics (Fink et al., 1984), the adjacent tidal inlets have received far less attention. During the early 1970's, Farrell (1970, 1972) studied coastal and estuarine processes in both the Saco and Scarborough Rivers and documented sedimentological and bedform trends on the tidal deltas, hydraulic conditions in the rivers, and the geologic history and shoreline changes of Saco Bay. Wave dynamics and longshore transport trends were investigated later at Wells Inlet in an attempt to solve shoaling problems in the harbor and entrance channel (Byrne and Zeigler, 1977; Bottin, 1978). In addition, Timson and Kale (1976), Nelson (1979), and Fink et al. (in press) determined shoreline changes at Wells Inlet prior to and following inlet stabilization. Finally, Trudeau (1979) and Nelson (1979) have looked at the effects of the Kennebec River and Morse River Inlet on the sedimentation history of the Popham-Seawall Beach system. The only other major studies on Maine's inlets prior to the 1980's were engineering projects conducted by the U.S. Army Corps of Engineers.

Beginning in 1982, the authors of this paper began a detailed investigation of different size inlets in Maine in order to determine the relationships among various inlet parameters and processes. This ongoing research has culminated in several publications (FitzGerald et al., 1984; Fink et al., 1984; Lincoln et al., 1985; FitzGerald and Fink, 1987; Barker, 1988; Lincoln and FitzGerald, 1988; Fink et al., in press) and is used as the data source for this paper. Our specific objectives are to describe the formation and morphology of different size inlets in Maine and to discuss the variability of their hydraulics and transport patterns. The tidal inlet terminology used in this paper follows that of Hayes (1975, 1979).

PHYSICAL SETTING

Bedrock Geology

The bedrock and structural geology of the coastal zone together with the extent and type of glacial deposits dictate the physiographic character of the shoreline and whether sediment was available for the development of coastal barriers and tidal inlets. Structurally, the coastal region is a litho-tectonic block, part of the Avalon terrain that was sutured to northeast America during the Devonian and subsequently intruded by batholiths of various compositions and ages (King, 1980; Osberg et al., 1985).

The complex geologic fabric of coastal Maine has been divided into four distinct physiographic compartments by Timson (1977). These divisions were later modified by Kelley and Timson (1983) and Kelley (1987). The divisions and the shoreline characteristics of the compartments are, as follows:

(1) Northeast compartment (Passamaquoddy to Machias Bays): Shoreline composed of high rocky cliffs formed on resistant metavolcanic rocks.

(2) North-central compartment (Machias to Penobscot Bays): Deeply embayed shoreline is characterized by resistant headlands and offshore islands formed from abundant granitic plutons.)

(3) South-central compartment (Penobscot to Casco Bays): Fluvial erosion accentuated by glacial scour has produced a peninsula coast with numerous embayments and deep estuaries formed from a high-grade metasedimentary fold belt that strikes in a northeast-southwest direction.

(4) Southwestern compartment (Cape Elizabeth to Piscataqua River): This section of coast has developed parallel to the regional structural trend and consists of bedrock headlands and intervening arcuate embayments formed from metasedimentary-metavolcanic units intruded by plutons.

Surficial Sediments

Repeated episodes of glacial excavation during the Pleistocene stripped the sediment cover from the coast of Maine leaving in its place a relatively thin till deposit (Thompson and Borns, 1985) which is somewhat thicker in bedrock valleys (Kelley et al., 1986). The till is commonly overlain by glaciomarine sediments (Presumpscot Formation), ranging in thickness up to 50 m and varying in composition from a pervasive fine-grained silty clay to coarser-grained silty sands that are commonly found toward the bottom or top of a sequence (Bloom, 1960; Thompson, 1987). The coarser units are thought to reflect higher energy conditions associated with currents flowing beneath the margin of the ice, as well as wave and fluvial reworking that was coincident with the regression of the sea, respectively (Thompson, 1987).

Southwestern Maine is particularly rich in sand due to the number of glaciomarine deltas that occur in this region and the high sand content of the Presumpscot Formation that borders most of the southern shoreline. This abundance of sandy deposits has assured a steady supply of sediments for the barriers of the arcuate embayment coast during Holocene sea level rise. Seismic studies in Saco Bay and elsewhere in the Gulf of Maine indicate that the glacial stratigraphy described above extends some distance offshore (Kelley et al., 1986; Belknap, 1987; Belknap et al., 1987b). Northeast of this zone the surficial cover in the coastal region is relatively thin, consisting of tills with frequent exposures of bedrock (Thompson and Borns, 1985), thereby reducing the availability of sand for resupply during the Holocene transgressive phase.

Hydrographic Regime

The highly variable orientation of Maine's shoreline and abundant offshore islands causes different exposures to the incident wave energy (Fink, 1978; Fink et al., 1984). Shoreline orientations coupled with a tidal range that doubles from its southern border to the Bay of Fundy produces a coastline with a wide range of wave and tidal energies.

Winds. Long-term wind data for Maine are available from Portland and Rockland airports, Brunswick Naval Air Station, Seguin Island Lighthouse, and Maine Yankee Atomic Power Company in Wiscasset (Nelson and Fink, 1978). A synoptic summary of meteorological observations is available for New England waters from NOAA (U.S. Naval Weather Service, 1976). An annual wind summary rose and a monthly compilation of recorded wind directions, velocities, and durations for a typical year from Seguin Island are shown in Figures 2 and 3, respectively (Fink et al., in press). The wind roses, sequential aerial photographs, and direct observations from a recording anemometer located at Popham Beach permit a determination of the dominant and prevailing winds as well as their seasonal patterns.

The prevailing winds blow from the western half of the compass with the months of May through September showing the greatest periods of calm while the winter months reflect strong wind activity from both extratropical storms and the passage of high pressure systems through the area. October and April are transitional months with strong east to northeast winds superimposed on either the increasing or decreasing intensity of the prevailing westerly pattern.

The strongest winds are from the northeast and are associated with extratropical storms that track east of the Gulf of Maine. Strong winds also blow from the northwest and southwesterly directions (Figs. 2 and 3). Because these winds are prevailing winds and at times are quite strong, they exert the greatest control on deviations from the astronomically induced tidal ranges. In turn, this influences tidal prism and tidal current conditions at inlets. The winds accompanying northeast storms not only affect tidal processes at inlets, but also significantly increase the transport of sand along the beaches to inlets.

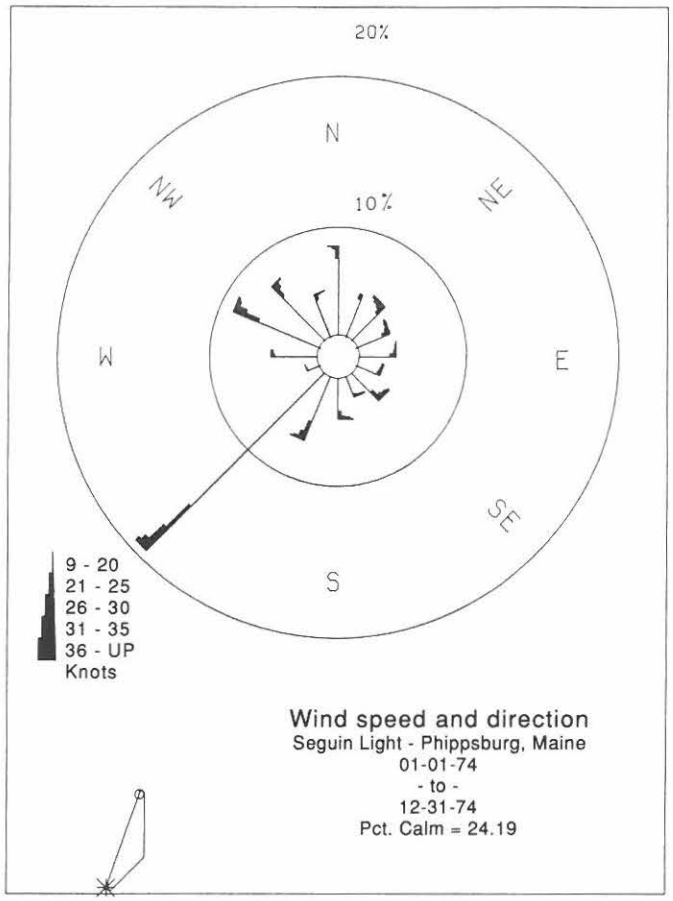

Figure 2. 1974 annual wind rose and resultant wind vector diagram for Seguin Island, Maine based on hourly readings (from Fink et al., in press).

Waves. Wave roses have been compiled by the U.S. Army Corps of Engineers for Penobscot Bay for deep-water conditions (Fig. 4; U.S. Army Corps of Engrs., 1957) and for 16 stations along the Maine coast for shallow water conditions (Jensen, 1983). These analyses were based on hindcast studies using three years and 20 years of wind data, respectively. The deep-water wave rose shows that waves over 1.5 m come predominantly from the east and northeast and that waves over 4 m come only from these directions, reflecting the influence of northeast storm winds over a long fetch. The average deep-water wave height at the Penobscot Bay site is 90 cm.

Farrell et al. (1971) summarized data for northeast storms and demonstrated that the average storm wave approach is 79°, having breaker heights between 0.9 and 1.4 m and periods of 6 to 8 seconds. They also showed that low pressure systems that track inland produce significant storm waves from the southeast with an average approach direction of 157.5°. Southeasters generally have lower wind velocities than northeast storms, have a shorter duration, and produce smaller storm surges, smaller wave heights, and shorter wave periods. Fink (1978) has plotted the distribution of wave approach directions for all the beaches along the coast of Maine in relation to northeast and southeast storms (Fig. 5). As seen in Figure 5, the orientation of most barrier and tidal inlet systems is such that they are susceptible to lower energy, less frequent southeast storms. Beach orientation in relation to northeast storms is, therefore, a significant factor in minimizing beach erosion, sand transport into inlets, destruction of ebb-tidal shoals, and other impacts of potentially high energy storms.

Seasonal information on the nearshore wave climate for the Maine coast (Jensen, 1983) in comparison to other U.S. coastal areas (Richardson, 1977) is shown in Figure 6, which is a summary of mean monthly nearshore wave heights. Jensen's (1983) analysis of wave energy for coastal Maine indicates that for the south-central and southwestern areas the average monthly wave height is greatest during the winter months (50 cm) and lowest during the summer period (34 cm). This region has slightly lower wave energy than the northern Atlantic coast and slightly greater energy than the Gulf of Mexico.

Tides. Tides in the Gulf of Maine are primarily driven by semidiurnal deep-water tides in the North Atlantic. At the edge of the continental shelf, the tide is 0.84 m (Fig. 7) and occurs

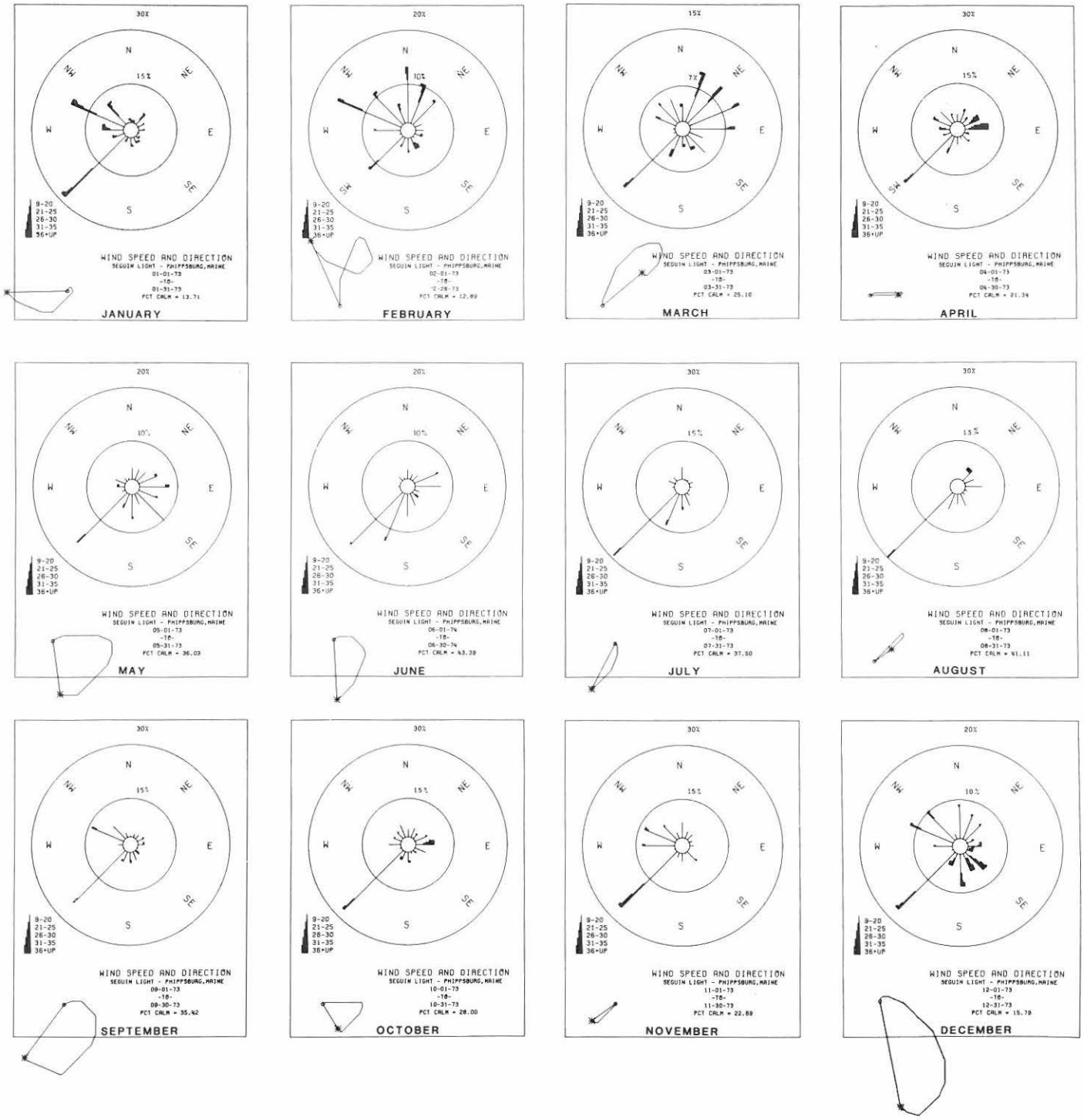

Figure 3. 1973 monthly summary wind roses and resultant wind directions for Seguin Island, Maine (from Fink et al., in press).

approximately 12 hrs after the moon's transit at Greenwich, England (Redfield, 1980). Tidal amplification, due to close correspondence between the lunar M_2 component and the natural resonant frequency of the Gulf of Maine (Garrett, 1972; Greenberg, 1979), increases the mean tidal range to 2.7 m along Maine's southern coast. During spring tide conditions, the range increases to 3.5 m. In the coastal classification scheme of Hayes (1979) and Nummedal and Fischer (1978), a tidal range of this

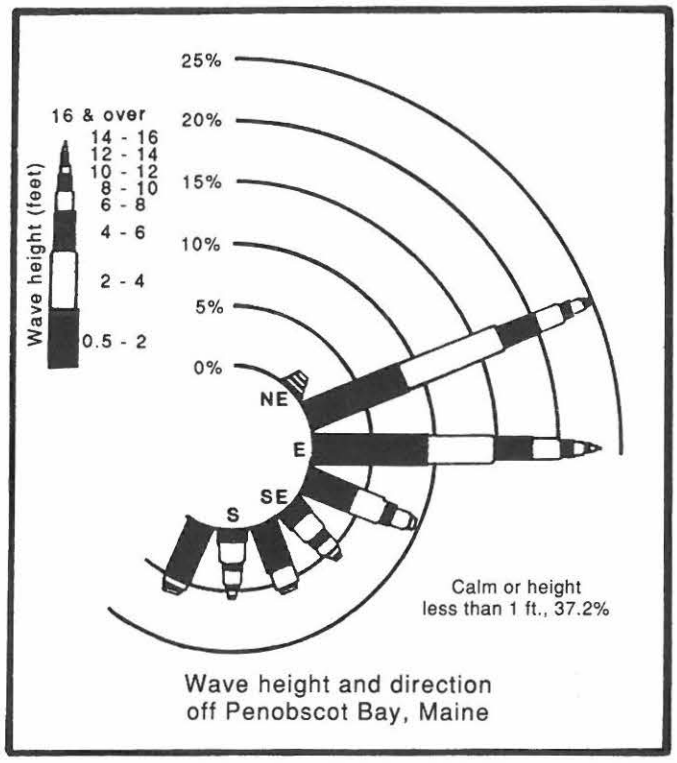

Figure 4. Wave diagram of deep-water wave heights hindcasted from three years of wind observations (1947-1950) (U.S. Army Corps of Engr., 1957). Dominant wave approach is from the east-northeast and is associated with northeast storms.

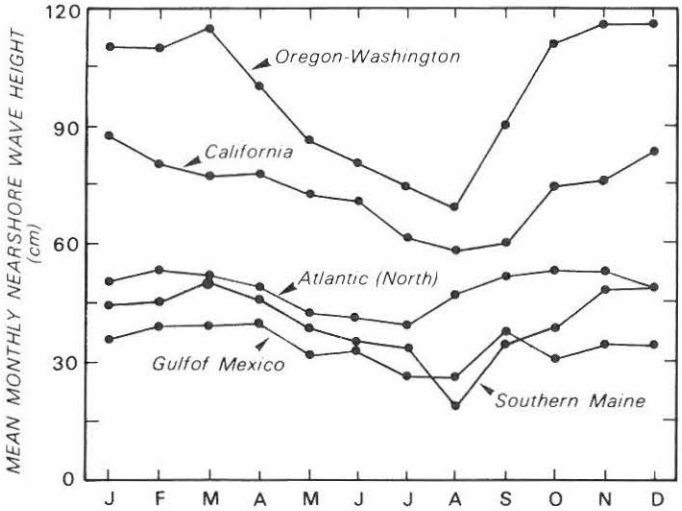

Figure 6. Monthly mean nearshore wave heights for various locations thoughout the United States (Richardson, 1977) including the south-central and southwestern coast of Maine (Jensen, 1983).

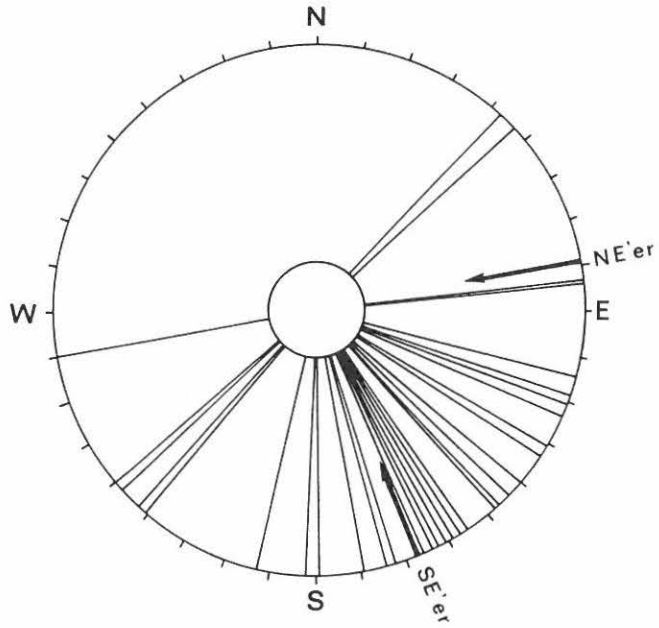

Figure 5. Distribution of wave approach directions for beaches in Maine with respect to northeast and southeast storms. The lines represent the wave approaches perpendicular to each barrier (from Fink, 1978).

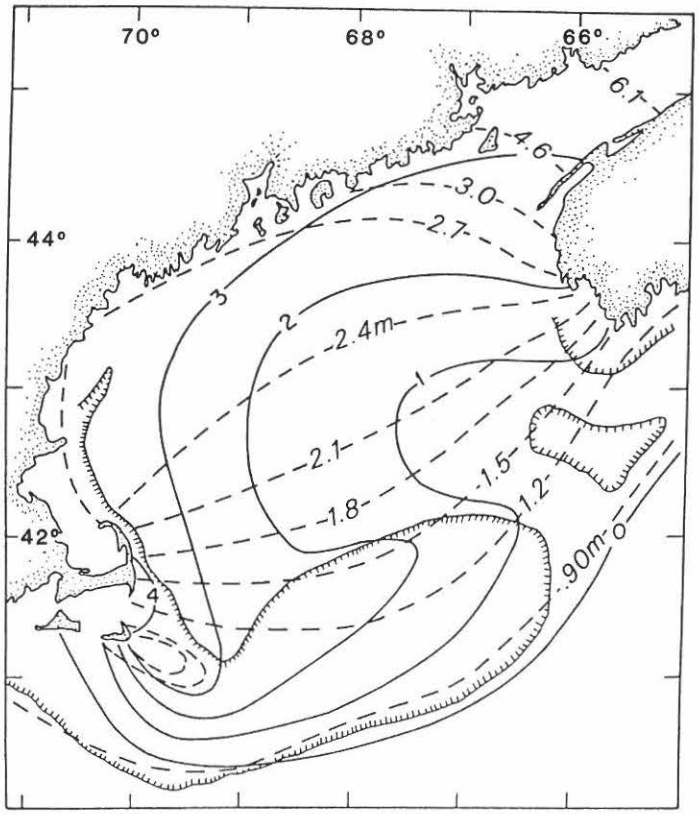

Figure 7. Corange-cotide lines for the Gulf of Maine amphidromic system. Lines of equal range are dashed and the lines of synchronous high tide (hours lag behind continental shelf tide) are solid (from Redfield, 1980). Hachures represent bathymetric highs.

magnitude and an average deep-water wave height of 90 cm places southern Maine inlets in a tide dominated-mixed energy regime (Fig. 8).

OCCURRENCE AND FORMATION OF TIDAL INLETS

Occurrence and Size Classes

The complex structural grain and highly variable regional sand supply along the coast of Maine have produced an uneven distribution of inlets along its shoreline; most of the 19 major inlets are found south of Cape Elizabeth. A classification of Maine's tidal inlets according to size is presented in Table 1. Channel width at the throat section is used as an approximation of inlet size, and generally as inlet width increases, so does channel depth. Three estuaries, the Saco, Kennebec and Sheepscot River mouths, are included in this list because they are morphodynamically associated with extensive beach systems and exhibit many of the morphologic characteristics and processes of tidal inlets. It can be seen (Table 1) that commensurate with the length of sandy barrier shoreline found there, most inlets are located in the southwestern (12) and south-central compartments (5), while the more concentrated, pocket barrier style of sand and gravel beaches in the sand-starved north-central and northeast compartments restrict the setting for inlet formation.

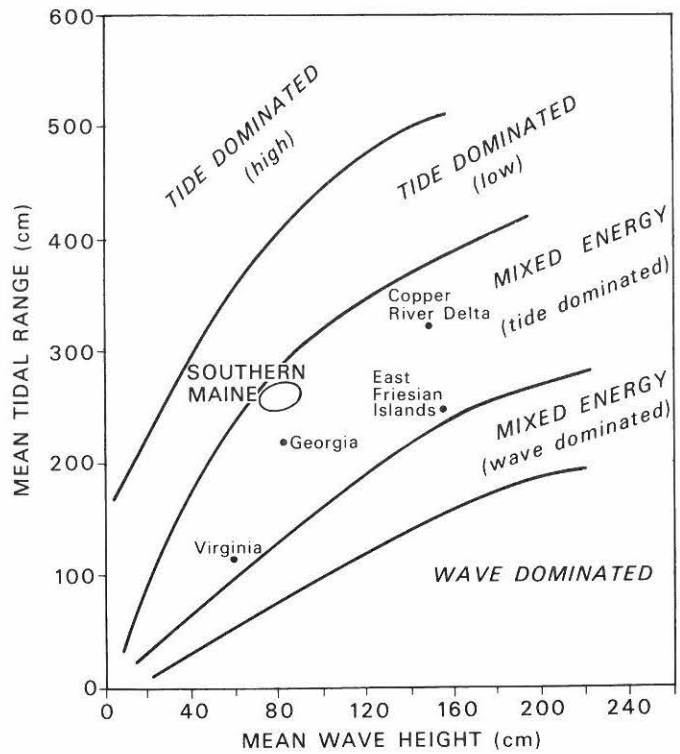

Figure 8. Shoreline classification based on tidal range and average deep-water wave height (Hayes, 1979; Nummedal and Fischer, 1978). Using this scheme, tidal inlets in the south-central and southwestern compartments of Maine plot in the mixed energy, tide-dominated setting.

TABLE 1. TIDAL INLET CLASSIFICATION

	Associated Barriers/Beaches	Location	Coastal Compartment
I. Estuaries			
A. Saco River	Hills Beach/Old Orchard Beach	Saco	Southwest
B. Kennebec River*	Popham Beach	Phippsburg	South-central
C. Sheepscot River	Reid Beaches	Georgetown	South-central
II. Large Inlets (width = 100-200 m)			
A. Scarborough River Inlet*	Surfside, and Western Beaches	Scarborough	Southwest
B. Kennebunk River Inlet	Kennebunk, and Goochs Beaches	Kennebunkport	Southwest
C. Wells Inlet*	Wells, and Drakes Is. Beaches	Wells	Southwest
III. Medium Inlets (width = 50-100 m)			
A. Ogunquit River Inlet*	Ogunquit-Moody Beach	Ogunquit	Southwest
B. Biddeford Pool Inlet	Hills Beach	Biddeford	Southwest
C. Spurwink River Inlet	Higgins Beach	Scarborough	Southwest
D. Mousam River Inlet	Parsons Beach	Kennebunk	Southwest
IV. Small Inlets (width below 50 m)			
A. Little River Inlet*	Crescent Surf/Laudholm Beach	Wells	Southwest
B. Batson River Inlet*	Goose Rocks Beach	Kennebunkport	Southwest
C. Little River Inlet*	Goose Rocks Beach	Kennebunkport	Southwest
D. Goosefare River Inlet	Ferry Beach/Old Orchard Beach	Saco	Southwest
E. Sprague River Inlet*	Seawall Beach	Phippsburg	South-central
F. Morse River Inlet*	Seawall, and Popham Beaches	Phippsburg	South-central
G. Little River Inlet	Reid Beaches	Georgetown	South-central
H. Sandy River Inlet	Sandy River Beach	Jonesport	North-central
I. W. Quoddy Head Inlet	South Lubec Beach	South Lubec	Northeast

*Inlets where field investigations have been performed.

The abundance of tidal inlets in the south-central and southwestern compartments is the result of an original sand source and a sufficiently steady resupply of sand, as well as an embayed coastal configuration and adjacent low-lying upland topography -- all factors that favor the formation and maintenance of extensive barrier systems. The abundance of sand-sized material is the result of glacially derived and riverine sediments that were reworked by waves and tidal currents along the transgressive Holocene sea level front. In contrast, coastal accumulations farther downeast developed from glacial deposits, mostly tills, that were poor in sand-sized material. The paucity of sand and proximity of high topographic relief limited the formation of sandy barriers essential to the development of tidal inlets. Instead, the coastal accumulations have taken on the form of fringing strandplains, tombolos, or pocket barriers with sediment sizes that are characterized by the coarser end members from pebbles to boulders.

The exact contribution of major rivers such as the Saco, Kennebec, and Sheepscot as original and continuous suppliers of sand to beaches is problematical. Today, only the Kennebec River may still be supplying sand to its associated barriers (Nelson and Fink, 1978; Nelson, 1979; FitzGerald and Fink, 1987), while no appreciable amounts of sand are introduced by the Saco (Farrell, 1970, 1972; Moreau et al., 1979) or Sheepscot Rivers. The beaches associated with these systems are likely the products of sand brought down the rivers during deglaciation. Belknap et al. (this volume) have identified a large paleodelta off the Kennebec River extending deeper than 65 m. During the Holocene transgression, it is likely that the top of this delta was reworked and a portion of this sand transported onshore where it became anchored to bedrock islands and headlands. Additional sand was probably added to the barrier systems directly from river-borne sediment load until modern dams reduced the supply. In the case of the Sheepscot River, its total sand resources and drainage area are far smaller than either the Kennebec or Saco, and it is likely that most of the sand comprising the Reid barriers came from the reworking of the Kennebec paleodelta.

Although most of the tidal inlets in Maine are referred to as "river inlets," the majority have very little freshwater discharge, especially when compared to their saltwater flow. For example, Scarborough River Inlet's freshwater discharge (3.1 m^3/s; Farrell, 1970) is four orders of magnitude less than its average saltwater flow (1.3×10^4 m^3/s; Barker, 1988). Field investigations have been conducted at 9 of the 19 tidal inlets in Maine (Fig. 1) and include at least one inlet in each of the four categories listed in Table 1. The morphologic and hydraulic characteristics of the inlets where studies were made are given in Table 2. Freshwater discharges listed in Table 2 were determined from rainfall and drainage basin data.

Formation

Barrier Construction. Radiocarbon dating of drowned tree stumps and backbarrier basal peats associated with existing barriers suggests that Maine barriers were formed between 6000 and 3000 years ago (Belknap et al., 1987a; Belknap et al., 1989; Bloom, 1960; Bradley, 1957; Hulmes, 1980; Hussey, 1959, 1970; McIntire and Morgan, 1963; Nelson and Fink, 1978). Although this time period is similar to that which has been determined for the East and Gulf Coasts (Ibrahim, 1986; Moslow and Colquhoun, 1981; Steele, 1980; Bernard et al., 1962), the Netherland mainland barriers (Van Straaten, 1965), the East Friesian Islands (Sindowski, 1973) and elsewhere, differences in the time frame of barrier development along the Maine coast can be expected due to its glacial history. The exact time at which the barriers and tidal inlets were sufficiently developed in Maine to prompt continuous backbarrier marsh accumulation depended upon the rate of sea-level rise and sand supply. For these reasons, it is likely that coastal barrier and tidal inlet systems in Maine had various times of origin and rates of development.

During the Holocene trangression, which began along the Maine coast approximately 8,000 to 10,000 yr B.P. (Belknap et al., 1987a), it is also possible that some barriers migrated to their present positions from locations offshore (Kelley et al., this volume). The tidal inlets along these transgressive barrier shorelines would have opened and closed and provided important transport pathways for the landward movement of sands.

At some point during the history of barrier construction in Maine, regressive barriers replaced some of the transgressive forms as the rate of sea-level rise slowed and the supply of sediment, primarily from offshore, increased. Hine et al. (1979) refer to this period as the "Late Holocene high sedimentation interval" and have explained the progradation of the North Carolina barriers as a consequence of this event. Along the coast of Maine, it is apparent that while some of these regressive barriers have persisted to the present, most barriers have become or remained transgressive through their long-term history (Nelson and Fink, 1978).

Tidal Inlet Evolution. The processes and controls involved in the evolution of barriers and tidal inlets in Maine are illustrated in a flow chart in Figure 9. In this scheme sediment supply, sea-level rise, and regional slope determine the time and location

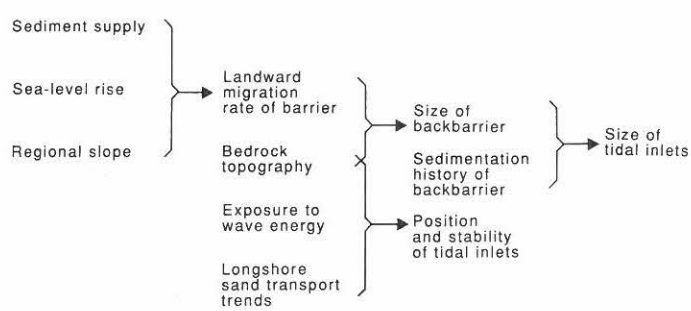

Figure 9. Flow chart of the evolutionary processes and controls in tidal inlet development for the Maine coast.

TABLE 2. MORPHOLOGIC AND HYDROLOGIC DATA

Inlet	Size Class	Physiographic Setting	Average Fresh Water Discharge (m^3/s)	Backbarrier Setting	Mean Tidal Range (m)	Tidal Currents (cm/s) Max Flood	Tidal Currents (cm/s) Max Ebb	Morphology Throat Sec. (m) Width	Morphology Throat Sec. (m) Depth	Morphology Sand Bodies Ebb Deltas	Morphology Sand Bodies Flood Deltas
Kennebec River Estuary	I	Glacially scoured bedrock valley	284.0	River (open water)	2.6	125	150	310	34.0	Subtidal, well developed	None
Scarborough River Inlet	II	Estuary, inlet stabilized by bedrock (one jetty)	3.1	Marsh & tidal creeks, open water & tidal flats at inlet mouth	2.7	144	165	150	5.0	Subtidal/ intertidal well developed	Well developed intertidal
Wells Inlet	II	Spit progradation (two jetties)	1.1	Marsh & tidal creeks, open water & tidal flats at mouth	2.6	70	110	130	5.8	Poorly developed subtidal	Well developed intertidal
Ogunquit River Inlet	III	Spit progradation across embayment, bedrock anchored	1.1	Marsh & tidal creeks, open water & tidal flats at mouth	2.7	109	98	70	2.4	Poorly developed subtidal	Multiple deltas, well developed
Little River Inlet (Kennebunkport)	IV	Spit progradation across embayment	1.0	Marsh & tidal creeks, small amount of open water & tidal flats at mouth	2.6	84	76	42	1.7	Poorly developed subtidal	Single delta, well developed
Batson River Inlet	IV	Beach ridge developed within embayment, bedrock anchored	0.5	Marsh & tidal creeks, small amount of open water & tidal flats at mouth	2.6	83	84	46	1.2	Absent	Multiple deltas, well developed
Little River Inlet (Goose Rocks)	IV	Beach ridge developed within embayment, bedrock anchored	0.3	Marsh & tidal creeks, small amount of open water & tidal flats at mouth	2.6	80	62	32	1.4	Poorly developed	Multiple deltas, well developed
Sprague River Inlet	IV	Spit progradation/ beach ridge developed, bedrock anchored	0.1	Marsh & tidal creeks, small amount of open water & tidal flats at mouth	2.6	92	75	26	1.2	Absent	Multiple deltas, well developed
Morse River Inlet	IV	Spit progradation/ beach ridge developed, bedrock anchored	0.2	Marsh & tidal creeks, drainstidal ponds, small amount of open water & tidal flats near mouth	2.6	121	59	42	3.0	Absent	Single delta, moderately developed

where barriers will form and the rate of their landward migration. In turn, the rate of barrier migration along with the bedrock topography of the region dictates the size of the backbarrier environment. The sedimentation history within the bay together with the area of the backbarrier control the tidal prism and the resulting size of the tidal inlets. The position of the inlet along the barrier is a function of bedrock topography, exposure to wave energy, and longshore sand transport trends.

In the scheme described above there are several evolutionary paths, two of which are depicted in Figures 10 and 11. In the first case, the barrier is exposed to the dominant east-northeast wave energy, and the ensuing longshore transport of sediment produces spit accretion across an embayment. This process continues until the size of the inlet equilibrates with the bay tidal prism. Normally, the inlet stabilizes against a bedrock headland at the downdrift end of the embayment. This morphology is evident at the Ogunquit River Inlet (Fig. 11a) where Moody and Ogunquit Beaches extend southward from Moody Point across a marsh-filled embayment, and the inlet channel is anchored next to bedrock. The north-south trend of the barrier results in sand transport toward the inlet during northeast storms.

In the second case, the embayment is protected from the dominant wave approach and thus, longshore transport is minimal (Fig. 10). Barrier construction under these conditions takes

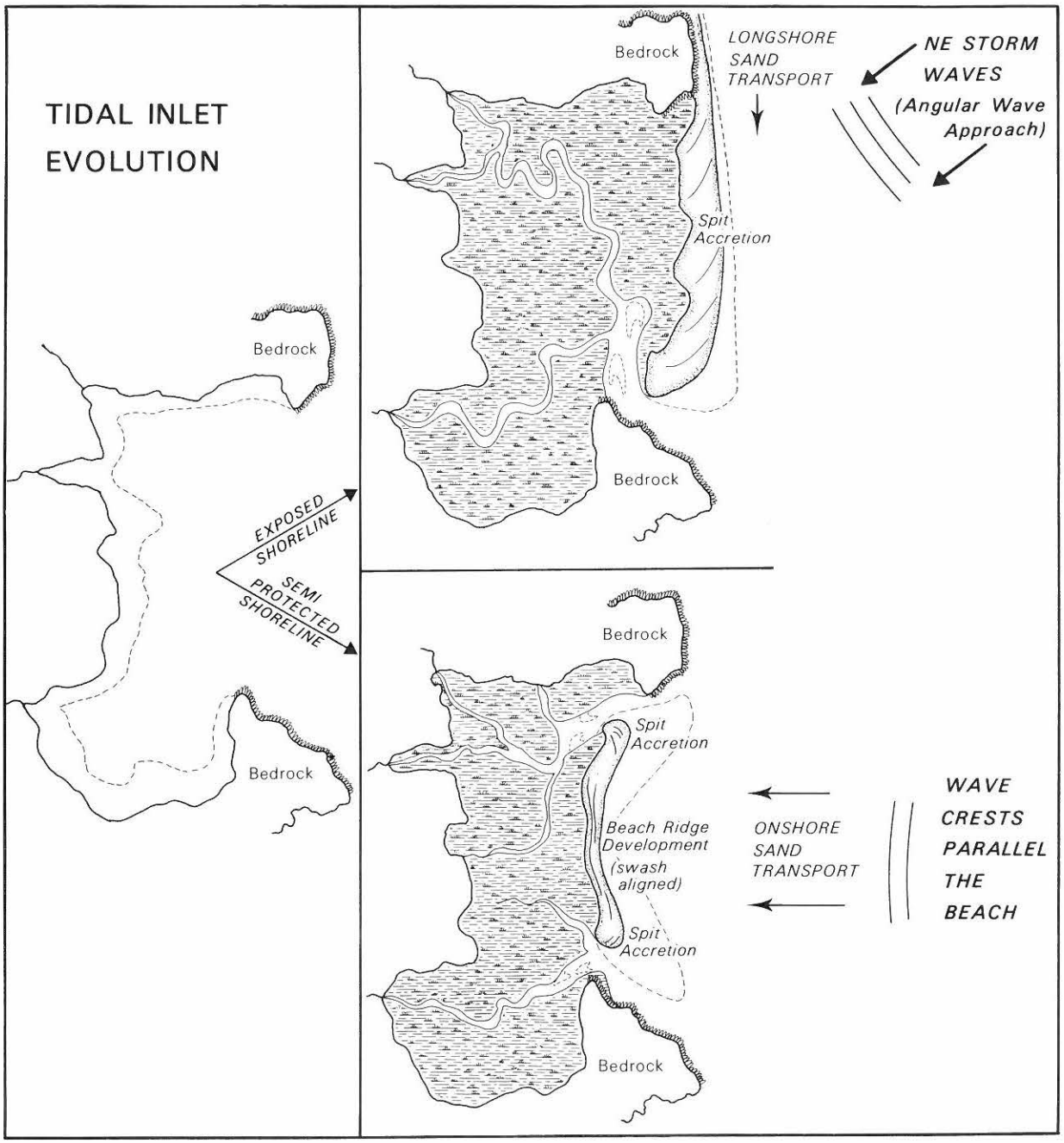

Figure 10. Conceptual diagram illustrating two mechanisms of inlet development along the Maine coast.

place through the onshore movement of sand which causes beach ridge formation and a progradation of the shoreline. Goose Rocks embayment exemplifies this situation well (Fig. 11b). This section of coast is sheltered from northeast storms by Timber Point and Timber Island (Fig. 1) and consequently, is most affected by south and southwest, low energy, constructional waves. Short period storm waves and wave refraction within the embayment may create local longshore transport resulting in elongation of the beach-ridge barrier. These conditions likely are responsible for the location of the inlets at each headland (Fig. 11b).

It should be emphasized that the two conditions discussed above are end members of the inlet evolutionary scheme depicted in Figure 10, and that the genesis of other barriers and inlets along the coast of Maine may incorporate processes of both the exposed and sheltered shoreline models. It should also be pointed out that

Figure 11. (a) 1974 oblique aerial photograph of Ogunquit River Inlet and Ogunquit Beach looking north (taken by Fink). (b) 1977 vertical aerial photograph of Goose Rocks Beach, Little River Inlet (top) and Batson River Inlets (bottom).

modification of the initial barrier form through its landward migration or sand redistribution may mask how the barriers and inlets originally developed.

The formation of the Sprague River Inlet-Seawall Beach complex along Maine's peninsula coast (Fig. 1) may be an example of an intermediate condition in the scheme shown in Figure 10. The drainage of the Sprague River Inlet is deflected to the west, which suggests that the barrier developed by spit progradation toward Cape Small. Presumably, this resulted from westerly longshore sand transport induced by northeast storms. Nelson (1979) also cites the presence of flood-tidal delta deposits, thin marsh peats, and curved beach ridges as evidence for a tidal inlet having been located in the middle of the barrier. If this is true, Sprague River Inlet may have migrated westward to its present position. In contrast to this evidence, ridge and swale morphology along the adjacent state park beach suggests that the barrier development in this region occurred through beach ridge accretion. This mode of formation is supported by present sand transport patterns and erosional-depositional trends. Thus, it appears that Seawall Beach may have developed from both the seaward progration of beach ridges and spit extension to the west. Through this process the Sprague River Inlet became anchored next to a bedrock headland.

INLET MORPHOLOGY

According to models of mixed-energy, tide-dominated coasts (Hayes, 1979; Nummedal and Fisher, 1978), Maine's tidal inlets should have well developed ebb- and flood-tidal deltas and be backed by marsh and tidal creeks with little open water area. However, the morphology of inlets along this coast exhibits a great deal of variability and generally does not contain all the features of mixed-energy inlets. These differences can be correlated with inlet size and can be explained in terms of tidal prism (Davis and Hayes, 1984), developmental history of the inlet, man-made influences, and sediment supply.

Small Inlets

Small inlets in Maine (width <50 m; Table 1) are characterized by shallow channels, single or multiple flood-tidal deltas (Fig. 12), and poorly formed or nonexistent ebb-tidal deltas. Most of these inlets are anchored next to bedrock headlands and bordered on the opposite side by a broad intertidal swash platform that merges with the low tide terrace of the adjacent beach. Seaward of the inlet throat, the main channel commonly migrates across the swash platform in response to changing wave climate.

Although the backbarrier of these inlets is relatively large, most of this region is filled with supratidal *Spartina patens* marsh. Hypsometric measurements of five small inlets in Maine by Lincoln and FitzGerald (1988) indicate that the backbarrier of these systems is comprised of 81% supratidal, 10% intertidal, and only 9% open water area. The extensive distribution of supratidal marsh has two important effects. First, because the marsh is at mean high water to a supratidal elevation, there is little change in bay area over the tidal cycle. As discussed in detail in a later section dealing with inlet hydraulics, this condition contributes to a flood dominance of these inlets, causing a net transport of

sand into their backbarriers. Secondly, the seemingly large backbarrier areas of these inlets actually produce small tidal prisms which result in small equilibrium inlet throat cross sections.

Medium-Sized Inlets

Medium-sized inlets (width = 50-100 m; Table 1) are similar in morphology to the small-sized inlets, but generally have deeper channels, larger and often multiple flood-tidal deltas, and more open water area and intertidal sand flats. The larger tidal prisms of medium-sized inlets produce better developed ebb-tidal deltas than those of the small-sized inlets. These deltas are mostly subtidal and their presence is evidenced by various ebb delta components including channel margin linear bars at Mousam and Spurwink Inlets and a terminal lobe off Biddeford Pool Inlet. The swash platform of these inlets, like Ogunquit inlet, is commonly cut by one or more shallow (depth <30 cm) flood channels.

The backbarrier of the Biddeford Pool Inlet is different from other inlets of this size class in that it comprises chiefly a sandy-mud tidal flat that is incised by a dendritic tidal creek system (Fig. 13). This large intertidal bay (2 km^2) formed behind two transgressive barriers that connected a large bedrock island to the mainland (Hulmes, 1980). Hulmes has demonstrated that the Pool was never an open water environment and is underlain by the Presumpscot Formation at a depth of 5 to 6 m. This unit is overlain by muds and sands that were transported through the inlet during flooding tides and through washover processes. Some sand material was also washed into the bay from the adjacent highlands (Hulmes, 1980). *Spartina* marsh has colonized the periphery of the bay and what has been interpreted by Hulmes (1980) to be a relict flood-tidal delta. The lack of an extensive marsh system behind the Biddeford Pool Inlet is probably due to slow sedimentation rates in the bay.

Figure 13. 1976 aerial photograph of Biddeford Pool Inlet and the backbarrier tidal flat region it drains. Note the peripheral marshes and small area of open water at low tide (taken by Fink).

Figure 12. (a) Aerial photograpgh of Batson River Inlets flood-tidal deltas (taken by Fink). (b) Ground photograph of Little River Inlet (Goose Rocks Beach) flood-tidal delta.

Large Inlets

Due to the harbors that Maine's large-sized tidal inlets (width = 100-200 m) provide, these inlets have undergone the greatest modification by man. The mouths to these inlets have been stabilized with jetties (Scarborough has a single jetty), and their main channels have had long histories of dredging. The bays inside the inlets have been cordoned off with numerous seawalls, bulkheads, and other revetments which have served to decrease the amount of backbarrier intertidal environment (Barker, 1988). Maintenance dredging and a substantial loss of intertidal area through man's encroachment have caused a disappearance of the flood-tidal delta at the Kennebunk River Inlet. Larger bay areas and perhaps a greater sand supply have maintained the presence of flood-tidal deltas at both Scarborough and Wells Inlets (Fig. 14).

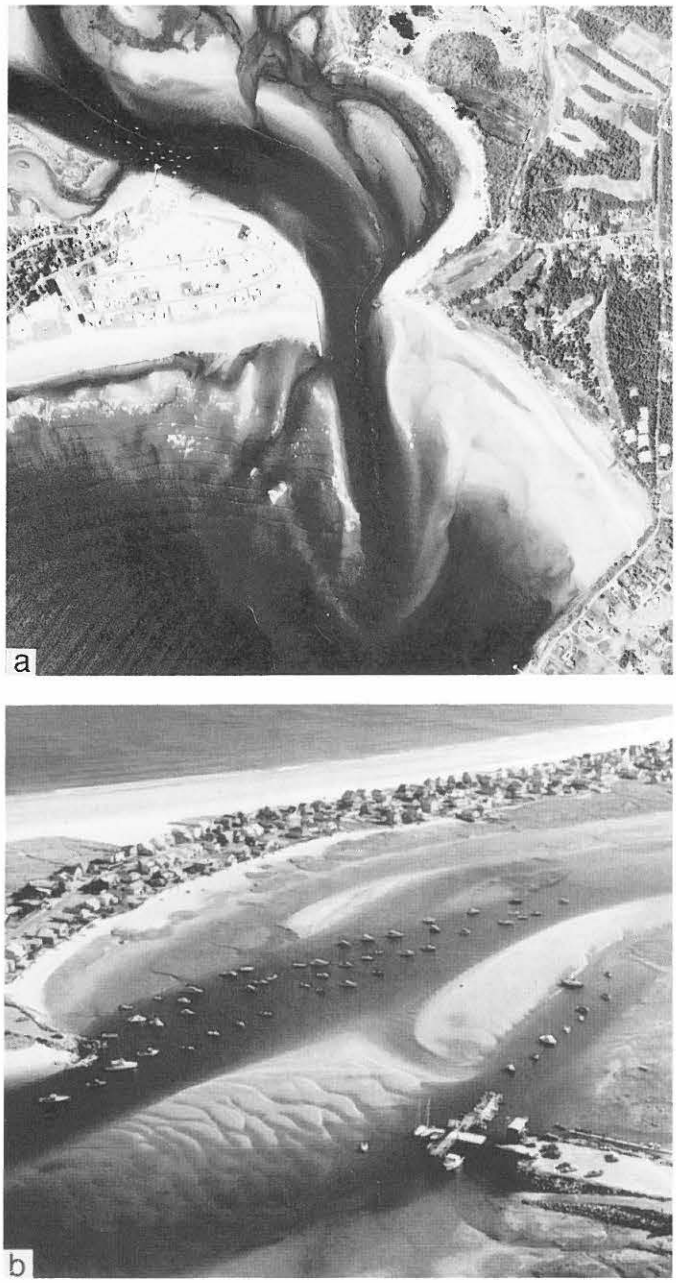

Figure 14. (a) 1970 vertical aerial photograph of Scarborough River inlet. (b) 1987 oblique aerial photograph of the Wells Inlet flood-tidal delta.

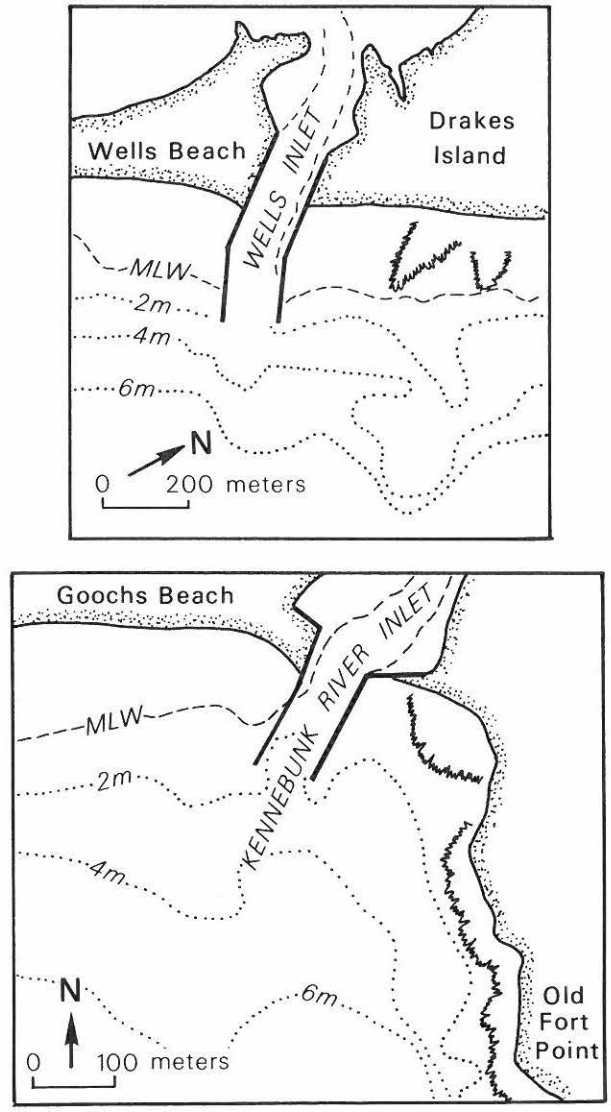

Figure 15. Bathymetric maps of Kennebunk River and Wells Inlets illustrating the lack of any significant ebb delta seaward of their jettied entrances (redrawn from National Ocean Survey chart #13286).

Although ebb-tidal deltas were present at both Wells and Kennebunk Inlets prior to jetty construction, the delta at Scarborough River Inlet (Fig. 14a) was considerably larger and far better developed. This condition continues to the present time and likely is due to the combination of the shallow region offshore of Pine Point and Western Beach, the low wave energy afforded by Prouts Neck, and an abundance of sand along this section of coast (Figs. 1 and 14a). All the components of Hayes' (1975) ebb delta model are clearly evident at Scarborough Inlet including a main ebb channel, terminal lobe, channel margin linear bars, marginal flood channels and swash bars. In contrast, the jettied entrances of the Kennebunk River and Wells Inlets exhibit little bathymetric evidence today of an ebb delta (Fig. 15). The length of their jetties (approx. 400 and 150 m, respectively) coupled with a smaller sand supply as compared to Scarborough Inlet has resulted in little sediment transport seaward of the jetties and/or most of the sand that is transported seaward being moved back onshore by wave action. The ebb deltas that are apparent in old aerial photographs of these regions prior to jetty construction were likely reworked by wave action and transported onshore. At both sites the adjacent beaches experienced accretion following completion of the jetty projects. It is possible that some of this accretion is attributable to erosion of the ebb deltas.

TABLE 3. ESTUARINE DISCHARGE

River	Freshwater Discharge Rate (m^3/s)	Freshwater Discharge Over 1/2 Tidal Cycle	Saltwater Tidal Prism (m^3)	Freshwater vs. Saltwater Discharge (Ratio)
1. Saco[a]	85	1.8×10^6	8.1×10^6	23%
2. Kennebec	283[b]	6.2×10^6	101.0×10^6 [c]	6%
3. Sheepscot	45[b]	1.0×10^6	76.0×10^6 [c]	1%

Note: a. Discharge and tidal prism data from Farrell (1970).
b. Discharge determined from precipitation and drainage area data.
c. Tidal prisms determined using the cubature method (Jarrett, 1976).

Estuaries

The Saco, Kennebec, and Sheepscot Estuaries (Table 1) are quite different from one another with respect to their morphology and sand transport patterns due to differences in sediment supplies, estuarine conditions (Table 3), and modifications by man. The morphology of the Saco Estuary has been drastically changed by jetties that were constructed in stages from 1868 to 1913 (Farrell, 1972; Nelson, 1979). Prior to this time, U.S. Coast and Geodetic Survey coastal charts (1866, #882; 1877, #6) indicate the presence of a large ebb-tidal delta which extended approximately 1.3 km from the mouth of the river seaward to the terminal lobe region where depths were less than 1.5 m at mean low water. After the jetties were in place, ebb discharge from the estuary no longer interacted with the littoral transport system and consequently, wave-generated sediment transport resulted in the onshore movement of the ebb delta sands. This caused temporary accretion along Camp Ellis Beach (Farrell, 1972). During the past century, sediment supply to the Saco estuary has been essentially cut off by several dams built upstream for the purpose of flood control and hydro-power. These dams, together with the jetty system, have reduced the Saco River's flood-tidal delta to a sub-tidal shoal (Farrell, 1972).

The significantly greater tidal prisms of the Kennebec and Sheepscot Estuaries as compared to the Saco Estuary (Table 3) are indicated by their larger size and the more important role that tides play in sedimentation processes. While both estuaries contain peripheral tidal flats and shallow upstream bays, channel depths inside the estuaries (depths >20 m) coupled with strong tidal currents preclude the formation of flood-tidal deltas. However, in Merrymeeting Bay at the confluence of the Androscoggin and Kennebec Rivers 26 km upstream of the river mouth, there are extensive sand bodies, some of which exhibit a flood-tidal delta morphology. Presumably, this shallow bay is a sediment sink that occasionally contributes sand to the downstream system during major flood events, like the 100^+-year flood of April 1987 (Fontaine, 1987).

Beaches at the mouths of both estuaries are asymmetrically distributed along their southwestern shorelines, while the opposite shoreline consists of exposed bedrock (Fig. 16). The preferential deposition of sediment along the western shore is undoubtedly the result of the dominant wave energy and the structural trend of this section of the Maine coast. During the Holocene transgression, it is probable that as the sands from the Kennebec glaciomarine delta (Belknap et al., 1987b) were reworked and transported onshore, they were also moved in a net westerly direction due to the dominant east-northeast storm waves (Figs. 4 and 5). Gradually this sediment accumulated against the northeast-southwest trending Phippsburg and Georgetown peninsulas. A sizable ebb-tidal delta, identified on coastal charts as the Pond Island Shoal, exists off the mouth of the Kennebec Estuary. This tidal delta, like the onshore beaches, is asymmetrically disposed about the mouth of the estuary and is entirely located on the western side of the main ebb channel. The lack of a corresponding ebb delta off the Sheepscot River is probably due to less sand in this system and the deeper channel system (Fig. 16). Depths in the main channel at the mouth of Sheepscot exceed 50 m as compared to the Kennebec where maximum depths are less than 15 m.

HISTORICAL SHORELINE CHANGES

Inlet Closure and Spit Breaching

During historical times, the inlets of Maine have undergone little net migration or change in position due to spit breaching (Nelson, 1979; Fink et al., 1984). The stability of the inlets is related to their location next to bedrock headlands and the fact that most of the inlets occur along swash-aligned beaches where there is little net longshore sediment transport to drive the inlet migration process. Furthermore, because the backbarriers of the inlets consist almost entirely of high-tide marsh, the breaching of a barrier is not a process that can occur easily. However, sedimentation and marsh development in the backbarrier area have led to the closure of at least one inlet in Maine and possibly others. In 1875 the Boston & Maine Railroad constructed an embankment across the backbarrier of Little River Inlet (Farrell, 1972). This decreased the drainage area of the marsh system, producing a smaller tidal prism which no longer was sufficient to keep Little River Inlet open. The relict inlet along Seawall Beach identified by Nelson (1979) and Fink et al. (in press) may be another example.

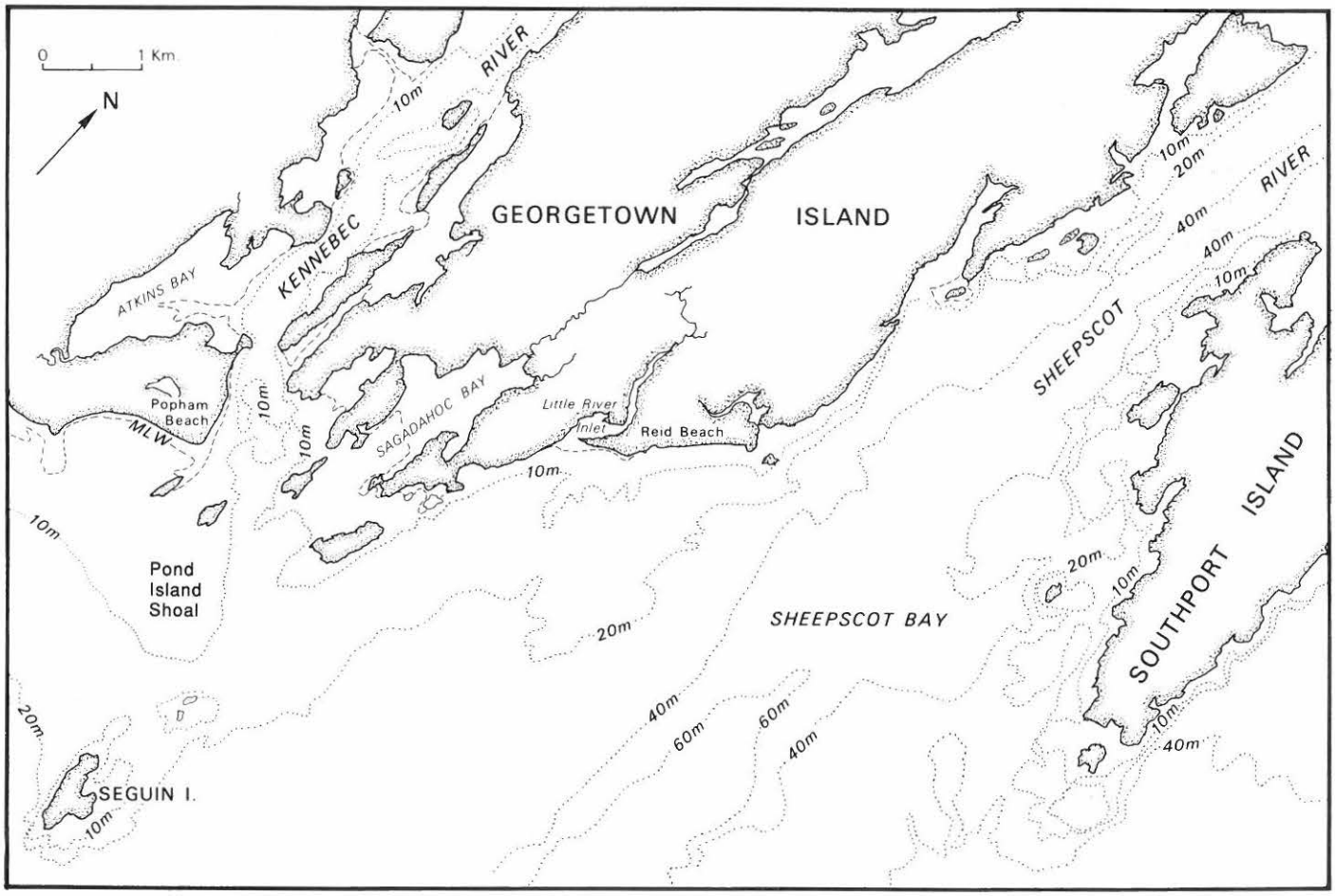

Figure 16. Map of the Kennebec and Sheepscot Estuaries (redrawn from National Ocean Survey chart #13293). Note the asymmetric distribution of beaches and ebb delta (at the Kennebec mouth) along the west sides of the estuaries. This development is likely a consequence of the dominant east-northeast wave energy of this coast.

To the authors' knowledge, during the recorded history of the Maine coast, the Ogunquit River Inlet is the only inlet that has changed positions (Fink et al., 1984; Fig. 17). A map dated about 1760 shows that the Ogunquit Inlet was once located about 1 km north of its postion today. The presence of a flood-tidal delta, which is now covered by a thin salt marsh deposit, at the same approximate site is geomorphic evidence that corroborates the inlet's 1760 location (Fink et al., 1984). The next historical map of this region, dated 1879, indicates that by this time the old inlet had closed and a new inlet had formed in its present position at the southern end of the beach (Nelson, 1979). Without further historical documentation or stratigraphic data, it can not be ascertained whether the relocation of the Ogunquit River Inlet was accomplished through inlet migration or spit breaching.

Inlet Associated Erosional-Depositional Processes

Although the throat positions of most inlets in Maine have remained essentially unchanged for at least the past 100 years, inlet processes have accounted for the greatest amount of shoreline change along the adjacent coast (Nelson, 1979; Fink et al., 1984). This is a characteristic that Maine inlets share with other mixed-energy inlets (FitzGerald et al., 1978). These erosional-depositional processes include the meandering of the main ebb channel, spit extension and retreat, storm overwash, and the transferral of sand between the ebb-tidal delta and landward beaches.

Main Ebb Channel Migrations. Migrations of the main ebb channel are best illustrated at the Morse River Inlet where repeated easterly excursions of the inlet channel have caused erosion of the adjacent Popham State Park beach (Fig. 18). This process has been described and illustrated with aerial photographs by Nelson (1979). The easterly migration of the channel is a result of the inlet throat being constricted by bedrock outcrops on both sides of the channel such that the ebb discharge is directed in an easterly course. As the channel migrates eastward, it lengthens, causing a decrease in the hydraulic slope and an increasingly inefficient exchange of water between the ocean and backbarrier. This situation is corrected when a new, shorter channel is cut through the westerly spit platform. The

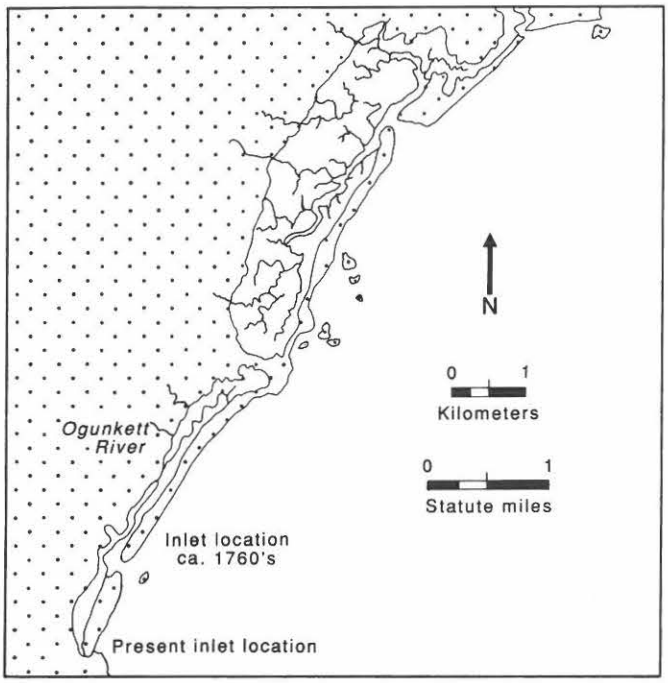

Figure 17. Early Colonial map showing former position of Ogunquit River Inlet (from Fink et al., 1984).

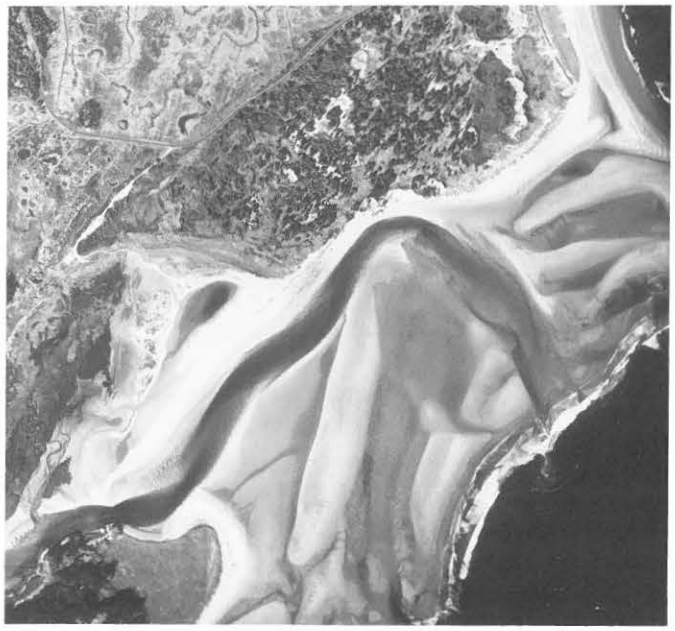

Figure 18. 1985 oblique aerial photograph of the Morse River Inlet depicting the migrational trend of its inlet channel.

breaching process, which normally occurs during major storms or high spring tides, has been documented with sequential maps that have been surveyed by Goldschmidt (1989) (Fig. 19). Easterly excursions of the Morse River Inlet channel have resulted in remobilization of sand that has been in long-term storage along the state park beach. The eroded sediment enters the littoral transport system and becomes part of a large sediment gyre that circulates sand along Popham Beach (FitzGerald and Fink, 1987).

Spit Accretion. Spit depositional and erosional processes appear to be the primary agent in modifying the mouths of most tidal inlets in Maine. The growth of recurved spits can cause a progradation of the ocean shoreline and/or constriction of the inlet. These processes are demonstrated well in the historical shoreline change maps of the southeast sides of the Spurwink and Scarborough River Inlets (Nelson, 1979; Fink et al., in press; Fig. 20). Even before the construction of the west jetty at Scarborough Inlet (completed in 1962), the Pine Point shoreline had a long history of easterly accretion and seaward progradation. Between 1914 and 1952 the shoreline built eastward about 400 m, while during the same period of time 100 to over 300 m of beach was added to the ocean shoreline. However, it should be noted that Nelson (1979) has identified earlier shoreline positions which indicate that Pine Point experienced substantial erosion from 1889 to 1914 and at several times prior to 1889 the shoreline extended to the same approximate position as it had in 1940 (Fig. 20a). Because of the inaccuracies of the older maps (Nelson, 1979), the real amount of erosion during this interval of time is unknown, as is the cause of the erosion.

As shown in Figure 20b, the history of northern Higgins Beach at Spurwink Inlet is similar to that of Pine Point, with one major difference. Since 1879, the Spurwink Inlet channel, like Scarborough Inlet, gradually narrowed (by more than 200 m) due to the northeastward extension of Higgins Beach spit. However, during the same period, the ocean-facing beach, unlike Pine Point lost over 50 m of shoreline. The differences in erosional-depositional processes at these two inlets can likely be explained by the greater abundance of sand along the Old Orchard-Surfside Beaches as compared to the Scarborough-Higgins Beach system. During the accretionary phase of the development of Pine Point, it is also possible that new sand supplies may have been introduced to the area as a consequence of jetty construction at the Saco Inlet and the ensuing onshore movement of its ebb-tidal delta sands. The closure of the Little River Inlet also freed sand that was stored in its ebb-tidal delta system (Farrell, 1972).

The interplay of sand resources between the ebb-tidal delta and the onshore beaches is an important process at large inlets and estuaries, particularly at the Popham Beach-Kennebec River Estuary system. This subject is discussed in the "Sediment Transport Trends" section later in this paper.

Jetty Construction. On a final note, it should be recognized that man's alteration of tidal inlets through jetty construction has accounted for large shoreline changes in Maine. The building of these structures has two primary effects. First, the longshore transport of sand becomes trapped on one or both sides of the jetties depending on their length and how they influence wave refraction patterns. Secondly, because flow from the inlet is confined by the jetties, the ebb jet extends further offshore. This

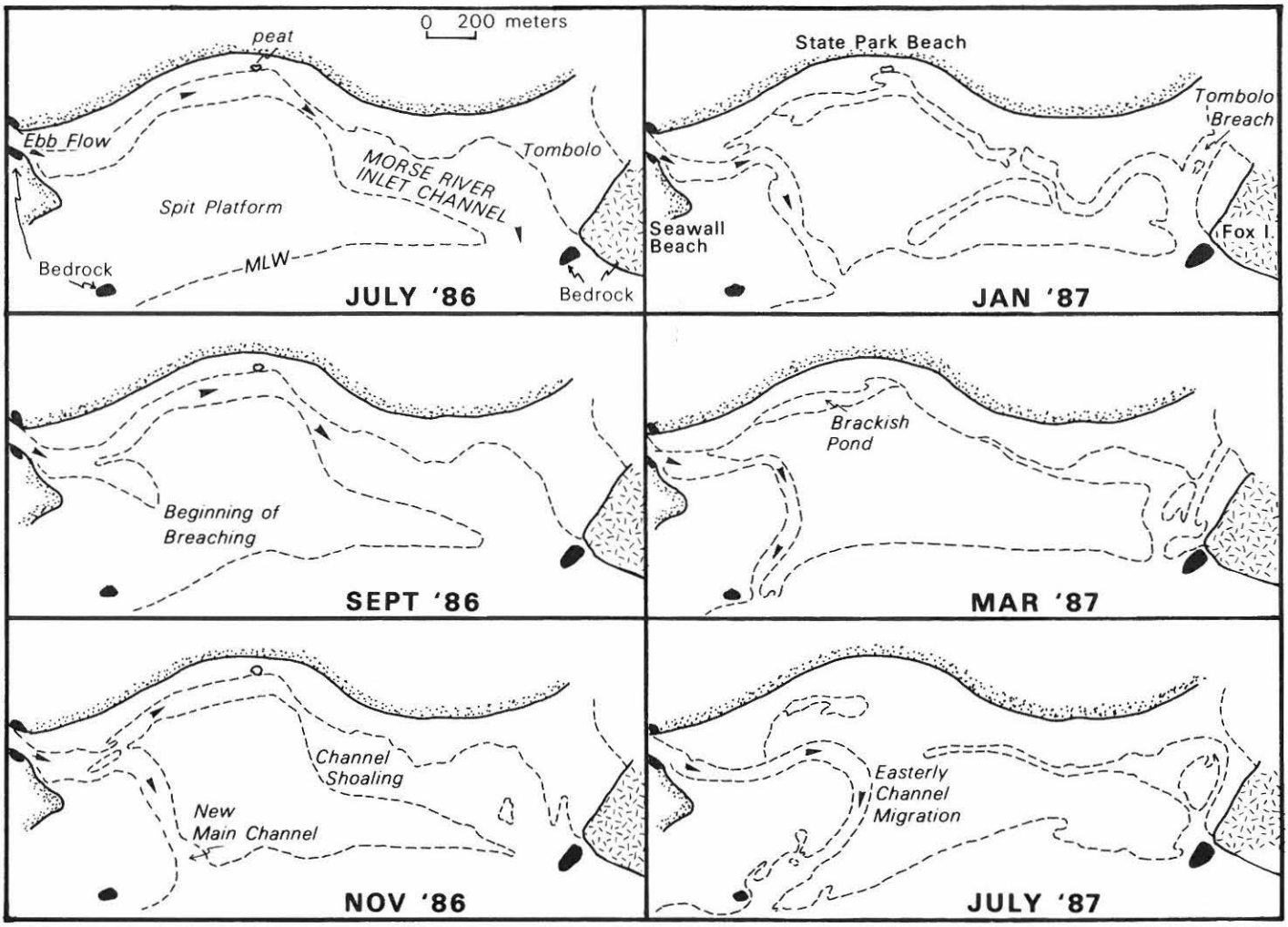

Figure 19. Sequential maps of the Morse River Inlet region covering a period when the inlet channel breached a new, hydraulically more efficient, pathway through the spit platform (from Goldschmidt, 1989).

results in a portion of the terminal lobe sands being transported into deeper waters while the rest of the ebb-tidal delta sediments are moved onshore by wave action. This latter process at Galveston Harbor accounted for 2.7×10^6 m^3 of sand moving onshore along the adjacent beaches (Morton, 1977). After jetties were constructed at Murrells Inlet, South Carolina, a detailed monitoring program by the U.S. Army Corps of Engineers documented the landward migration and attachment of a large swash bar to the beach south of the inlet (Douglas, 1987). As mentioned earlier, this process has also occurred at the Saco River Estuary during the early 1900's (Farrell, 1972).

Jetty construction at Wells Inlet in 1962 has led to a 200 m seaward migration of the shoreline on both sides of the inlet (Timson and Kale, 1976; Nelson, 1979; Fink et al., in press; Fig. 21). These jetties have been extremely effective sediment traps due to their lengths and the bidirectional longshore sand transport pattern that characterizes this section of coast (Byrne and Zeigler, 1977). The jetties extend past the high tide shoreline by over 200 m and thus, intercept most of the littoral sand moving along the coast. While angular wave approach causes sand deposition along the updrift side of the inlet, the downdrift shoreline is sheltered from wave action by the jetties and therefore, loses little sediment. Wave refraction around the jetties also promotes stability along the immediate downdrift shoreline. The impoundment of sand next to the jetties has seriously depleted the sand supplies of the adjacent beaches, resulting in erosion of Wells Beach and Drakes Island, and to a lesser degree Laudholm Beaches (Fink et al., 1984).

TIDAL INLET HYDRAULICS

Tidal forcing has a significant influence on the dynamics of tidal inlets in Maine and these controls vary with inlet size. To examine these influences, we present current velocity time series data (Table 2) and the results of tidal harmonic analyses for the medium-sized Ogunquit River Inlet and the large-sized Scar-

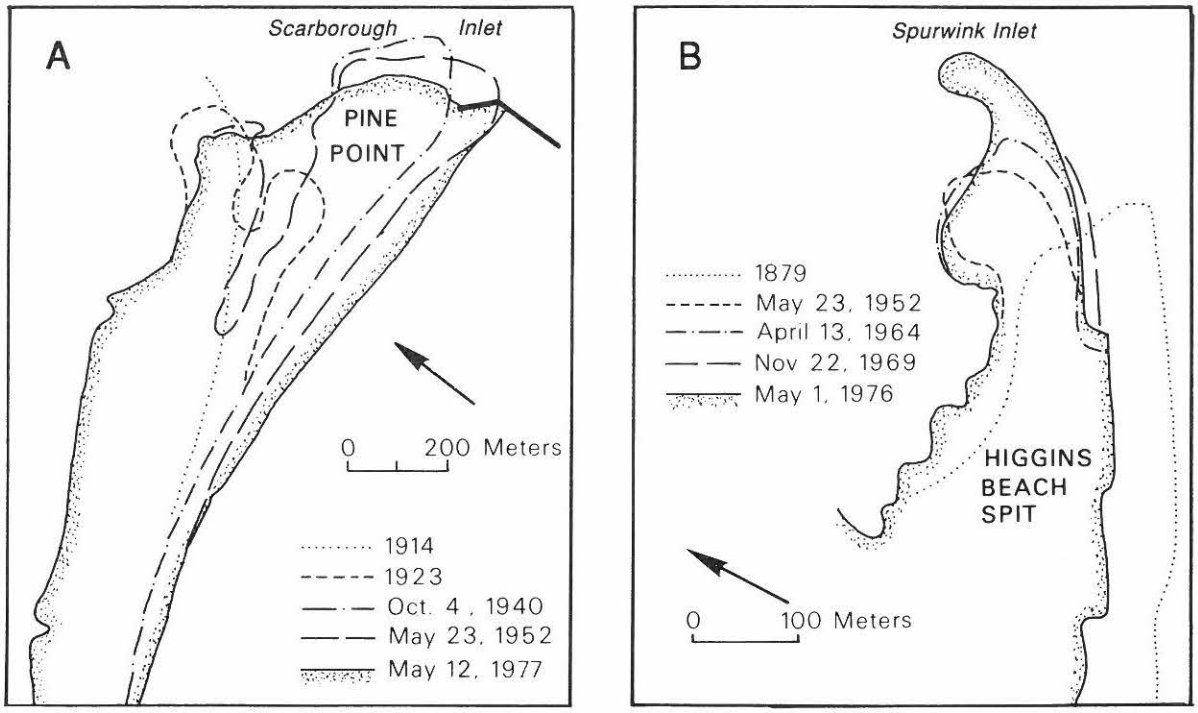

Figure 20. Historical shoreline change of (a) Pine Point and (b) Higgins Beach spit (after Nelson, 1979).

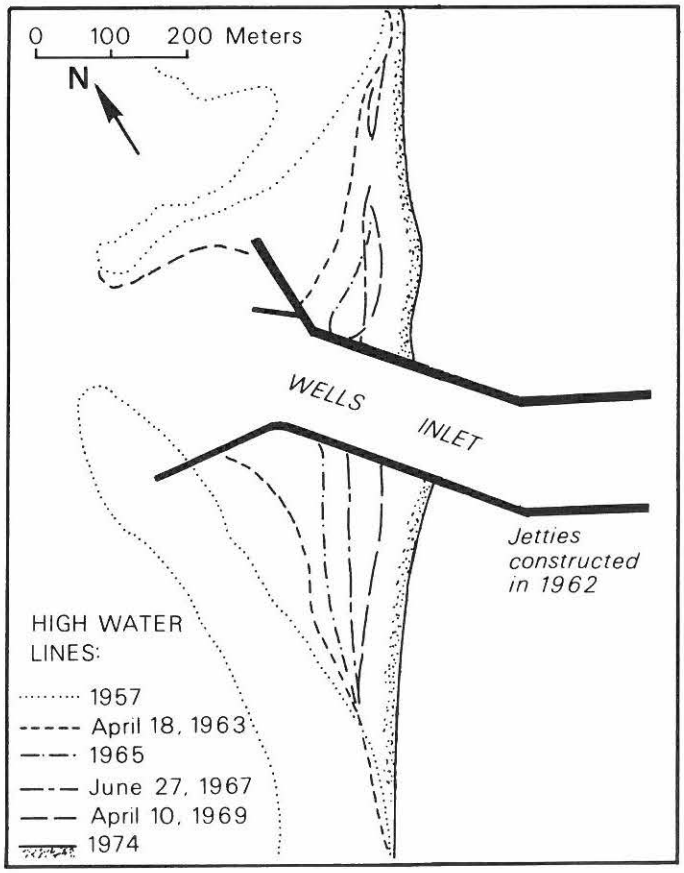

Figure 21. Shoreline changes at Wells Inlet (after Nelson, 1979).

TABLE 4. TIDAL CONSTANTS FOR OGUNQUIT OCEAN TIDE

		Speed (deg/hr)	Amplitude (meters)	GMT[1] Phase (deg)
Principal lunar	M_2	28.9841	1.293	101.82
Principal solar	S_2	30.0000	0.184	139.36
Larger lunar elliptic	N_2	28.4397	0.304	58.91
Luni-solar diurnal	K_1	15.0410	0.132	194.62
M_2 overtide[2]	M_4	57.9682	0.012	336.48
Principal lunar diurnal	O_1	13.9430	0.112	180.96
M_2 overtide	M_6	86.9523	0.007	125.84
S_2 overtide	S_4	60.0000	0.002	79.03
S_2 overtide	S_6	90.0000	0.002	129.76
M_2 overtide	M_8	115.9364	0.001	313.69

[1]Greenwich Mean Time
[2]Higher frequency multiple of the astronomic tidal component

borough River Inlet. Although a thorough discussion of estuarine processes is beyond the scope of this paper, some tidal current data are presented for the mouth of the Kennebec River in the "Sediment Transport Trends" section.

The Forcing Tide

Amplitude and Greenwich mean time phase (Greenwich epoch (G) in Schureman, 1958) were calculated for a 29 day record from a site 200 m offshore of the Ogunquit River Inlet using the least squares harmonic analysis of Boon and Kiley (1978) (Table 4). The form number ($Nf = (K_1+O_1)/(M_2+S_2)$) at

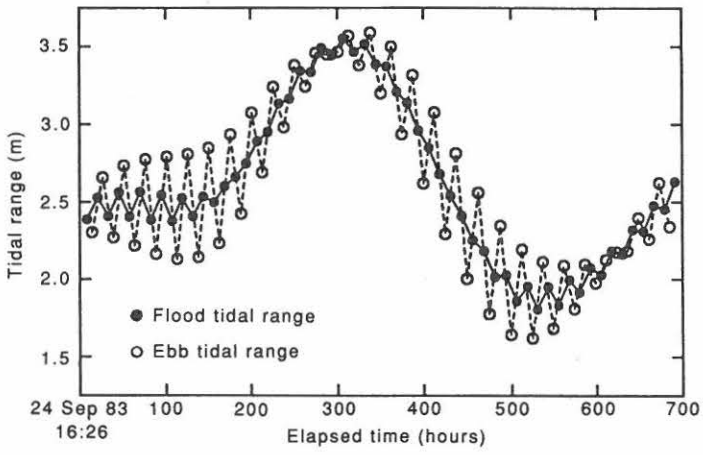

Figure 22. Plot of ebb and flood ranges for a 29 day tide record from 200 m offshore of the Ogunquit River Inlet. The diurnal inequality is almost totally accounted for in the ebb half of the tidal cycle which is alternately larger than and then smaller than the flood range (from Lincoln and FitzGerald, 1988).

this location equals 0.17, indicating a strong semidiurnal tide (Defant, 1958). The primary forcing constituent is the lunar semidiurnal M_2 with an amplitude of 1.29 m. The mean tidal range for the 29 day tide record is 2.62 m with a maximum spring range of 3.68 m and a minimum neap range of 1.66 m. Average diurnal inequality is 36 cm. Tide constants were similar for a 29 day tide record from a site 200 m seaward of the Scarborough River Inlet. A plot of tidal ranges for the 29 day period demonstrates that the ebb range of the ocean tide is alternately greater than and then less than the flood range (Fig. 22). Assuming a constant inlet cross section and bay area, stronger ebb currents are required to pass the alternately larger ebb tidal prism through the inlet. Likewise, when the ebb tidal range is alternately smaller than the flood, ebb currents should be weaker than the flood currents. However, this particular tidal range-tidal current relationship is not observed within the small inlets along Maine's coast, indicating significant modification of the tide as it propagates through the inlet channel.

Small and Medium-Sized Inlets

Tide and Current Asymmetry. Tide durations measured for full tidal cycles (flood and ebb) at Morse, Sprague, Little (Goose Rocks), Batson, and Ogunquit River Inlets (Fig. 1) indicate that the ebb exceeds the flood by an average of 3:22 hours (range: 1:45 to 5:15 hours, Table 5). The tide curve at these small inlets is characterized by a steeply rising tide followed by a less steeply falling tide with an extended period of slack water (Fig. 23). The following flood cycle begins abruptly. Low water occurs later in the inlet and bay than for the ocean tide, whereas the time of high water is nearly synchronous with the forcing ocean tide. At the same inlet throat stations, mean flood current velocity (depth and time-averaged) exceeded those of the ebb by an average of 21 cm sec^{-1} while maximum flood currents (depth-averaged) are 20 cm sec^{-1} stronger than those of the flood (Table 5). Assumimg no significant difference between ebb and flood tidal prisms (i.e. small diurnal inequality and small fresh water influx) and assuming no significant change in the average channel cross-sectional area between ebb and flood, a simple conservation of mass argument can be used to explain the observed flood current dominance: because there is a shorter period of time to fill the backbarrier than to empty it, flood current velocities must exceed ebb current velocities.

Shallow Water Effects on Inlet and Bay Tide. As the ocean tide passes through the shallow channel of small and medium-sized inlets in Maine, two tide characteristics vary systematically:

(1) Amplitude decay due to friction occurs in all semidiurnal components (M_2, S_2, N_2; Fig. 24) and,

(2) Asymmetry between shorter flood and longer ebb durations increases (Fig. 25; Table 6).

TABLE 5. INLET THROAT TIDE CURRENTS AT FIVE SMALL MAINE INLETS

		Maximum Velocity		Average Velocity		Duration		Tidal Range	
Location	Date	F(cm/sec)	E(cm/sec)	F(cm/sec)	E(cm/sec)	F(hrs)	E(hrs)	F(cm)	E(cm)
Morse	6/19/82	155	117	121	59	4:24	7:52	183	173
	6/26/82	200	148	102	44	3:44	8:59	174	191
Sprague	7/14/82	92	75	47	32	3:52	8:14	147	123
	8/04/82	91	90	51	33	3:18	8:06	147	121
Little	7/27/82	80	62	45	43	5:10	7:20	238	228
Batson	8/02/82	83	84	38	39	4:50	6:35	194	184
Ogunquit	8/18/82	109	98	58	48	4:08	7:36	204	206
	8/19/82	110	78	62	36	4:32	7:40	214	220
	1/28/83	83	58	38	34	4:12	7:06	248	242
	1/29/83	120	68	68	35	5:42	8:03	297	290

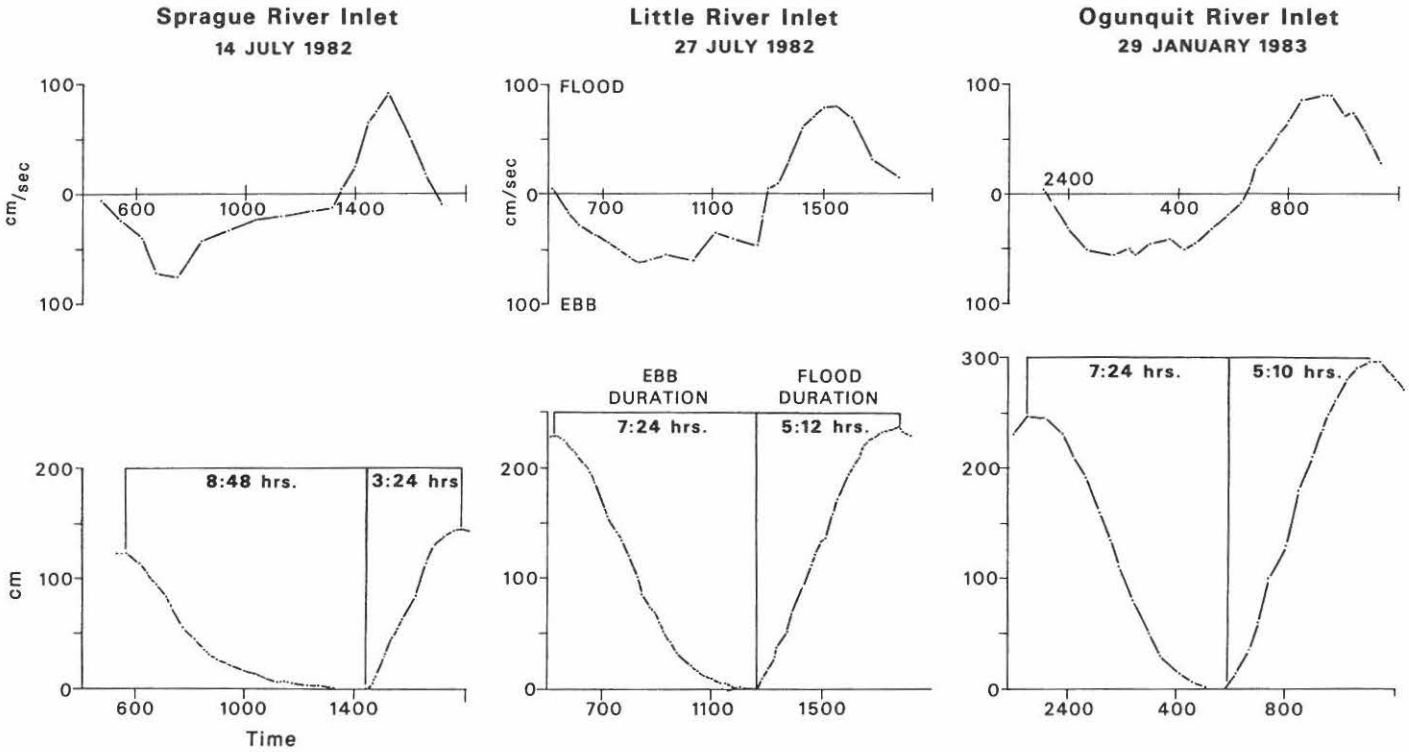

Figure 23. Representative plots of tide levels and velocity time series measured at the throat section of three small-sized tidal inlets. Maximum and mean flood currents are stronger than those for the ebb. Ebb durations exceed those of the flood.

There is no significant variation in the diurnal K_1, and O_1, components, apparently a common observation for shallow inlets (Aubrey and Speer, 1985). Some semidiurnal energy is transferred to higher frequency tidal components called overtides (see Table 4), rather than lost to friction (Fig. 24). Boon and Byrne (1981) and Aubrey and Speer (1985) have shown that progressive development of tidal asymmetries within inlets and estuaries along the U.S. Atlantic coast may be satisfactorily modeled by the growth of the M_4 overtide relative to the M_2. The ratio M_4/M_2 varies with the magnitude of the asymmetry while the sense of asymmetry is reflected by the relative phase of the components.

Pingree and Maddox (1978) isolated advection, friction, and continuity (conservation of mass) terms in the equation describing fluid flow through an inlet as the primary sources of non-linear harmonic growth of overtides. In small, constricted channels, advection is negligible. Numerical models of inlet/bay systems, having backbarriers that are filled with supratidal marsh and tidal creeks with steep banks, predict prolonged ebb durations and flood dominance (Mota Oliveira, 1970; Seelig and Sorenson, 1978; Speer and Aubrey, 1985). The supratidal marsh occupies about 81% of the total backbarrier area at small to medium-sized inlets in Maine; the rest is 10% intertidal sand shoals and 9% subtidal channels and open bay area (Lincoln and FitzGerald, 1988). Marsh areas are only inundated during the highest spring tides and periods of storm surge. Once the intertidal sand shoals are overtopped by the rising tide (elevation <0.5 m), the bay areas of the inlets do not vary significantly. A non-linear continuity argument does not explain harmonic growth of overtides at these small to medium-sized inlets, although it does appear to corroborate the modeling efforts mentioned above. Seelig and Sorenson (1978) predicted flood current dominance in shallow channels where friction increases significantly as a function of decreasing water depth. High friction at small inlets in Maine, which is evidenced by a 38% attenuation of the tidal amplitude over a distance of only 2.3 km at the Ogunquit River Inlet, is likely to be a major cause of overtide (M_4) generation.

In addition to the factors noted above, Lincoln and FitzGerald (1988) proposed that flood current dominance (prolonged ebb duration) may be caused by a truncation of the lower portion of the ocean tide by the very shallow channels of these inlets (often 30 cm shallower than the ocean mean low water). Tide curves, which were reconstructed from the M_2 and M_4 tidal components using the three Ogunquit tide records shown in Figure 25, illustrate this truncation: low tide is raised more by attenuation plus truncation at the throat and bay stations than high tide is lowered by attentuation alone. As the ocean tide falls below the level of the channel, the position of low tide is controlled by channel depth (Fig. 26). Semidiurnal inequalities and fortnightly variations occur only in the high tide positions

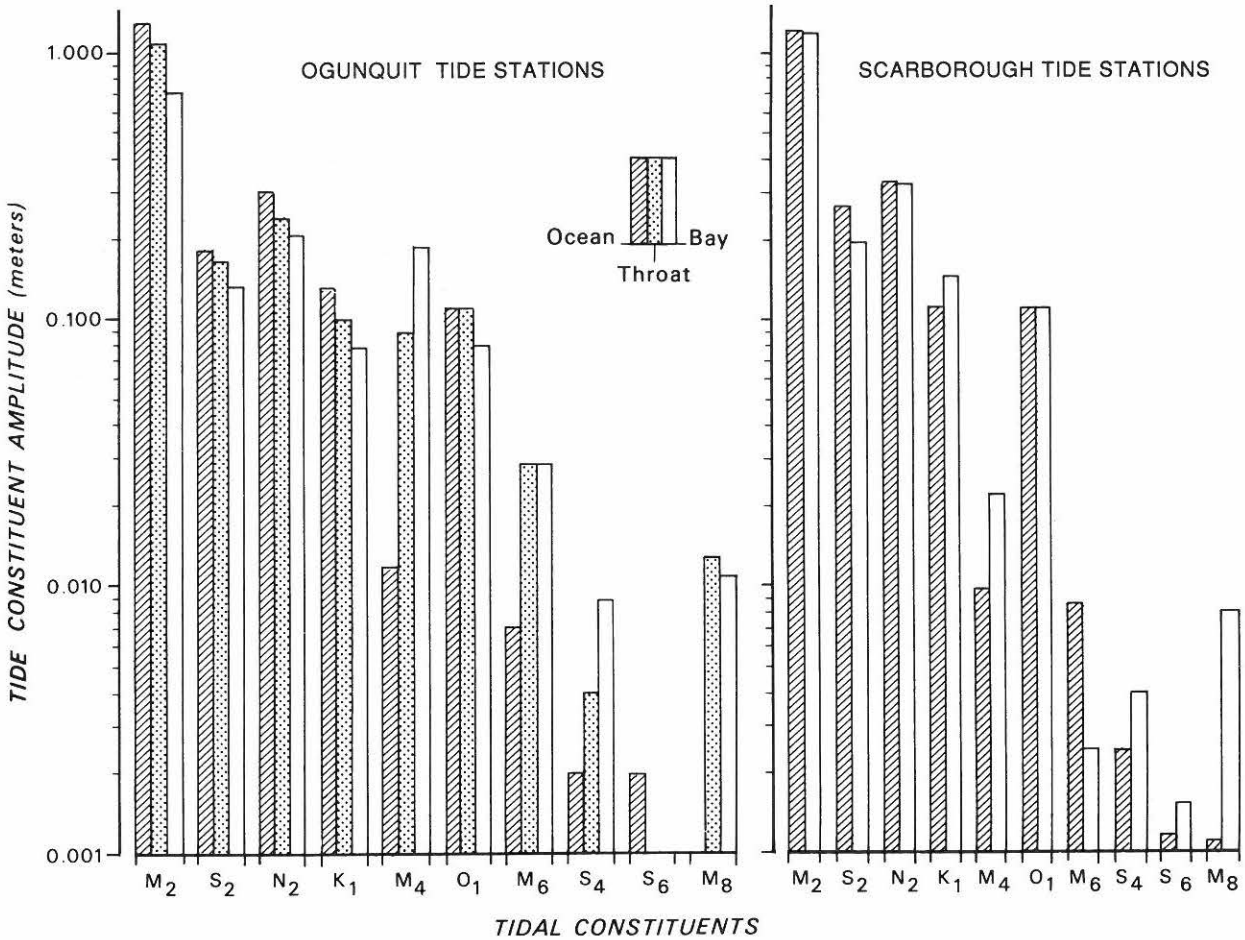

Figure 24. Bar graph of 10 tidal constituents for three stations at the Ogunquit River Inlet: 200 m offshore, throat section, and in the bay, and two stations at Scarborough River Inlet: 200 m offshore and in the bay.

since the minimum low tide level cannot fall below the channel floor. Inefficiency in emptying the dense marsh and tidal creek system in the backbarrier prevents the small tidal inlet channels from running dry during low tide.

Tide and Current Asymmetry at Large-Sized Inlets

Tide records from the seaward station at the large-sized Scarborough River Inlet (Table 1) show tide range and duration asymmetry similar to the Ogunquit ocean tide station (Tables 2 and 5). As should be expected, constants for the forcing tide are nearly the same for both inlets since they are only separated by about 30 km (Fig. 24). However, as the tide propagates through the deeper and wider Scarborough Inlet channel, flood duration increases slightly at the inlet throat and does not begin to exhibit flood dominant shallow water effects until the tide reaches the bay station (Fig. 24, Table 6).

Compared to Ogunquit Inlet, attenuation of diurnal components at Scarborough Inlet is minimal, indicating less friction in the channel at large-sized inlets. Growth of overtides is also small, indicating that non-linear advection and continuity effects do not influence the inlet tide significantly. Due to a slight asymmetry which favors longer flood durations at the Scar-

TABLE 6. ATTENUATION AND ASYMMETRY OF THE TIDE AT A SMALL AND MID-SIZED INLET

	OGUNQUIT RIVER INLET		
	Ocean[1]	Throat[1]	Bay[1]
Tide Range (meters)	2.62	2.28	1.62
Ebb Duration (hours)	6.17	6.75	7.73
Flood Duration (hours)	6.25	5.68	4.49
	SCARBOROUGH RIVER INLET		
	Ocean[1]	Throat[2]	Bay[1]
Tide Range (meters)	2.73	2.98	2.59
Ebb Duration (hours)	6.17	6.00	6.35
Flood Duration (hours)	6.25	6.40	6.07

[1]Average of 29 day tide record
[2]Average of 4 full tide cycle records

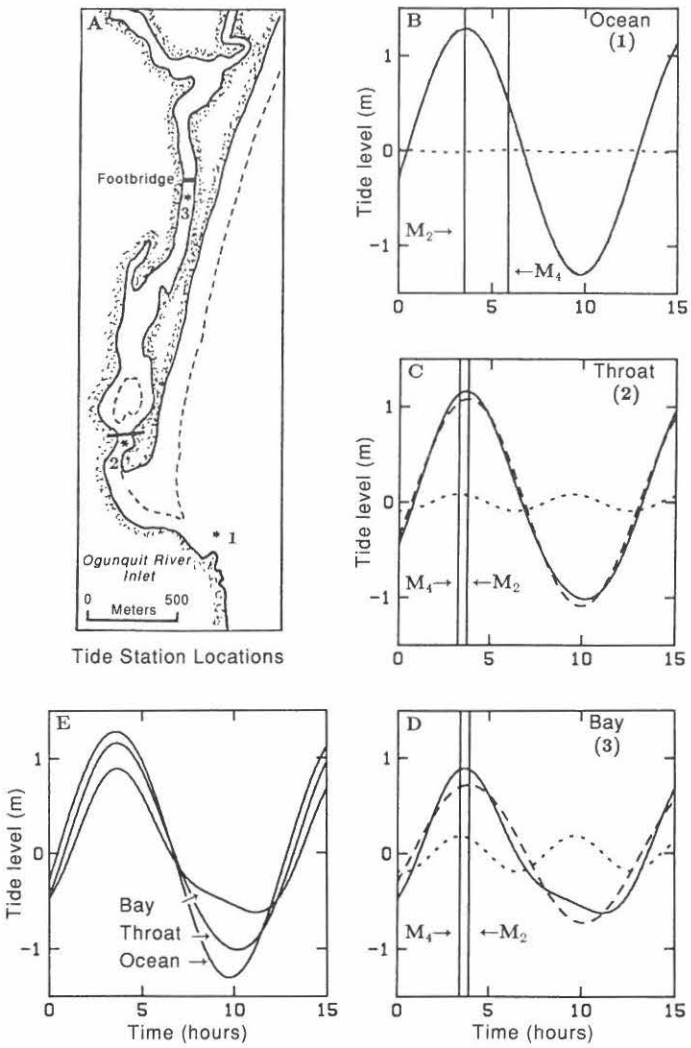

Figure 25. (a) Tide record locations at Ogunquit River Inlet. (b,c,d) Overlay of M_2 (dashed line), M_4 (dotted line) and combined $M_2 + M_4$ (solid line) tides for each record. Vertical lines illustrate tha phase relationship between the M_2 astronomic component and its overtide (M_4). (e) Overlay of combined tides from (b-d) shows the progressive attenuation of the tidal amplitude, near synchronous times of high-tide among the stations, and increasing low-tide lag time towards the bay station (from Lincoln and FitzGerald, 1988).

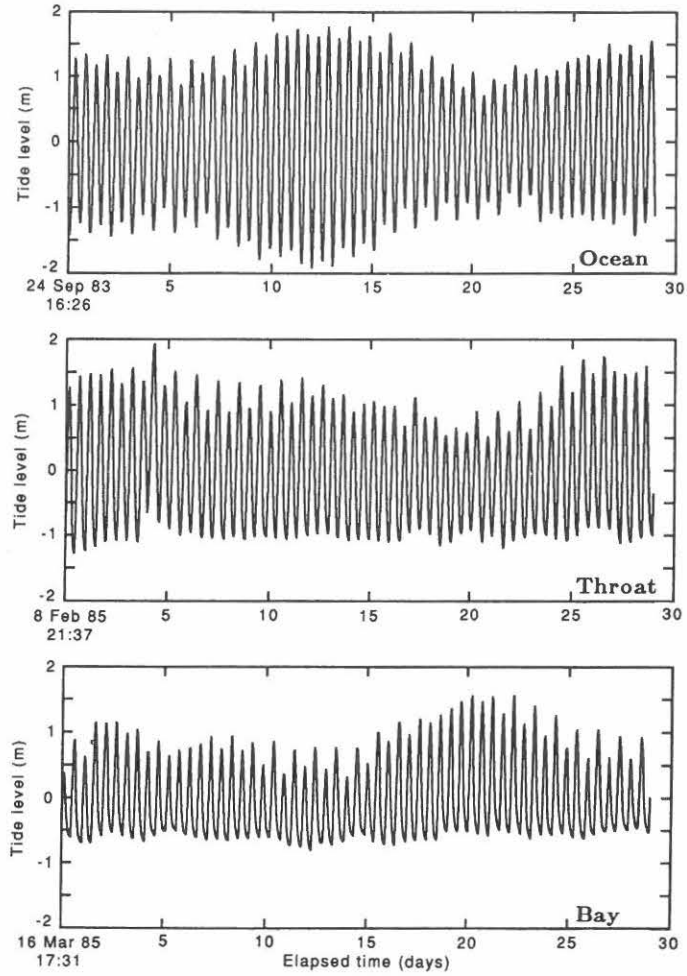

Figure 26. Complete tide records from three Ogunquit Inlet locations (Fig. 25). Note that bathymetry controls the low tide level. Diurnal inequalities and fortnightly variations are also eliminated from low tide as the channel shoals above the ocean mean low water level (from Lincoln and FitzGerald, 1988).

borough inlet throat, the same continuity argument that was used to explain the flood current dominance at small and medium-sized inlets can also be employed to predict the ebb dominance at Scarborough Inlet. For any given tidal range at Scarborough Inlet, both maximum and mean ebb current velocities exceed that of the flood at the inlet throat (Barker, 1988; Fig. 27).

The backbarrier at the Scarborough River Inlet has more intertidal area and a smaller percentage of high marsh than at small and medium-sized inlets (Barker, 1988). Non-linear basin filling due to large intertidal areas within the backbarrier (i.e. tidal flats and low marsh) was shown to be the primary cause of ebb dominance at Wachapregue Inlet, Virginia (Boon and Byrne, 1981) and at Price Inlet, South Carolina (FitzGerald and Nummedal, 1983). FitzGerald and Nummedal (1983) argued that a large bay area at high tide causes inefficient inlet filling, whereas a small bay area at low tide allows more efficient exchange of water. Low efficiency at high tide causes the time of high water in the bay to lag behind the ocean tide, whereas lag time for low tide is minimal. This is opposite to the situation observed at small and medium-sized inlets where high tide is nearly synchronous with the ocean tide and time of low water lags far behind the ocean tide. As noted above, Lincoln and FitzGerald (1988) proposed that low water lag at small and medium-sized inlets is due to the time required for the ocean tide to rise above the channel bottom.

These fundamental differences between the way the ocean tide interacts with small and medium-sized versus large-sized

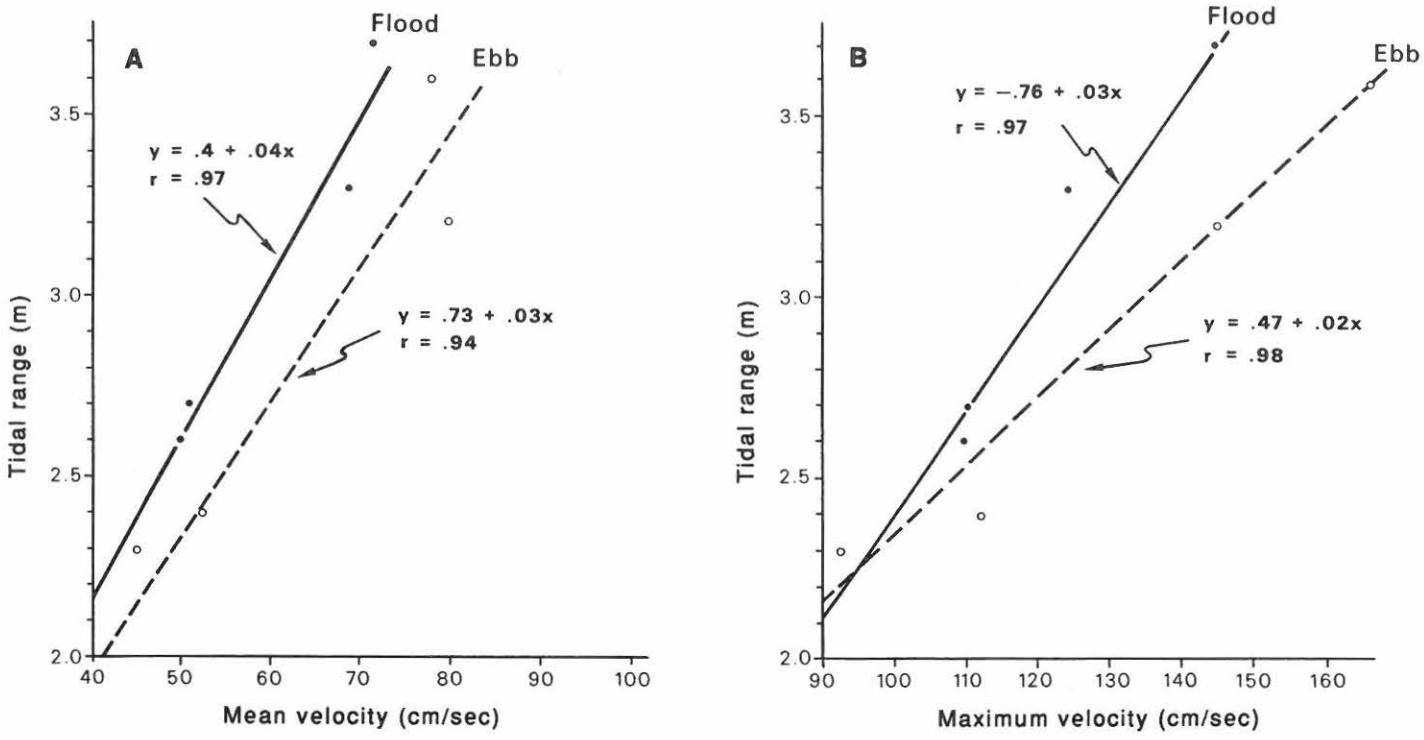

Figure 27. Plots of tidal range versus: (a) mean current velocity and (b) maximum current velocity at the Scarborough Inlet throat (from Barker, 1988).

inlet channels control tidal asymmetry and thus profoundly affect the sedimentation patterns at these types of inlets.

SEDIMENT TRANSPORT TRENDS

In this section, sand transport trends are briefly described for the Ogunquit and Scarborough River Inlets and at the mouth of the Kennebec River. While the sand transport patterns at the Ogunquit River Inlet are quite similar to dispersion processes at other small to medium-sized inlets, sedimentation processes at Scarborough Inlet and the Kennebec Estuary are less representative of their size classes (Table 2). This is because many of the large inlets have jettied entrances and/or vastly different sediment abundances. More detailed discussions of the these inlets can be found in FitzGerald et al. (1984), FitzGerald and Fink (1987), Lincoln and FitzGerald (1988), and Barker (1988).

Ogunquit River Inlet

While longshore transport rates are generally quite low along the Maine coast (<10,000 m^3/yr; Byrne and Zeigler, 1977; Fink et al., 1984), northeast storms are capable of moving large quantities of sediment along the shore during a relatively short period. Along Ogunquit Beach a net southerly transport direction, which would be predicted by the dominant northeasterly waves, has been documented by wave process measurements and sedimentological data (Lincoln et al., 1985). Once sand reaches the vicinity of the inlet, wave-generated and flood-tidal currents transport the sediment across the spit platform and into the inlet channel (Fig. 28). Sand in the outer portion of the main channel is moved in a net seaward direction by the dominant ebb tidal currents, eventually being deposited on the ebb delta. Here, wave action moves the sand back onshore to the spit platform and the adjacent beach, thus completing the counter-clockwise sediment gyre.

Sand which is deposited in the inlet channel along the landward portion of the spit platform is transported into the backbarrier by dominant flood-tidal currents (FitzGerald et al., 1984; Fig. 28). This sediment accumulates on flood-tidal deltas, in channels, and in other intertidal environments. This landward movement of sand has been corroborated with geomorphic evidence. In 1974, the flood delta was mined of its sand for dune construction along the adjacent Ogunquit Beach (Fig. 29a). One year later, the flood-tidal delta had completely reformed to original size and configuration (Fig. 29b).

Scarborough River Inlet

The similarity of the 1970 vertical aerial photograph of the Scarborough River Inlet in Figure 14 to the present condition indicates that little has changed at the inlet over the past 17 years. A detailed historical study of the inlet by Barker (1988) covering

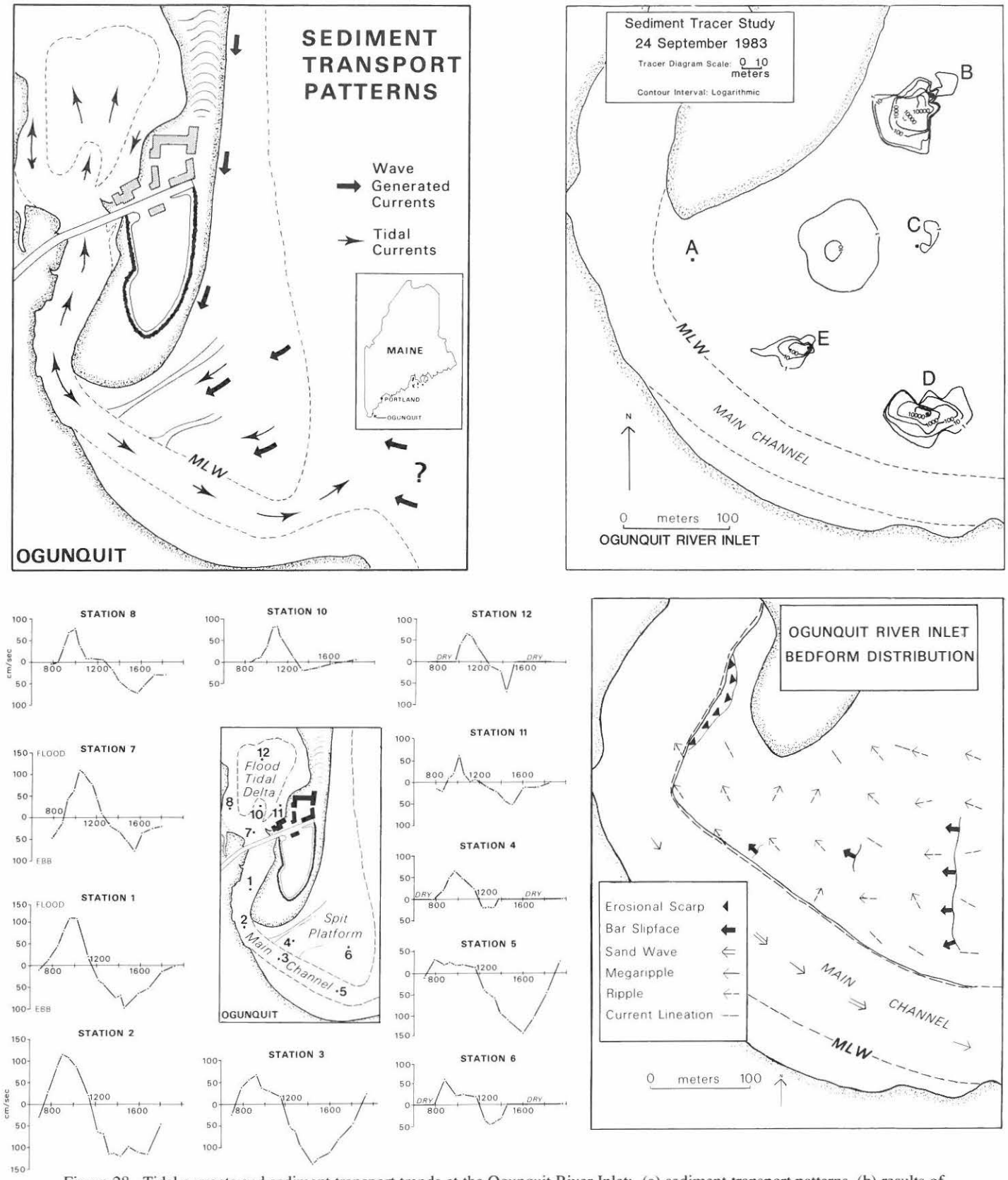

Figure 28. Tidal currents and sediment transport trends at the Ogunquit River Inlet: (a) sediment transport patterns, (b) results of sediment tracer experiments, (c) tidal current time series (stations 1-6, 18 August 1982; stations 7-12, 19 August 1982), and (d) bedform migrational trends, 6 October 1981 (from Lincoln et al., 1985).

Figure 29. Aerial photographs of the Ogunquit River Inlet showing the flood-tidal delta (a) after it had been mined of its sediment (1974), and (b) one year later (1975) after it had completely reformed (taken by Fink).

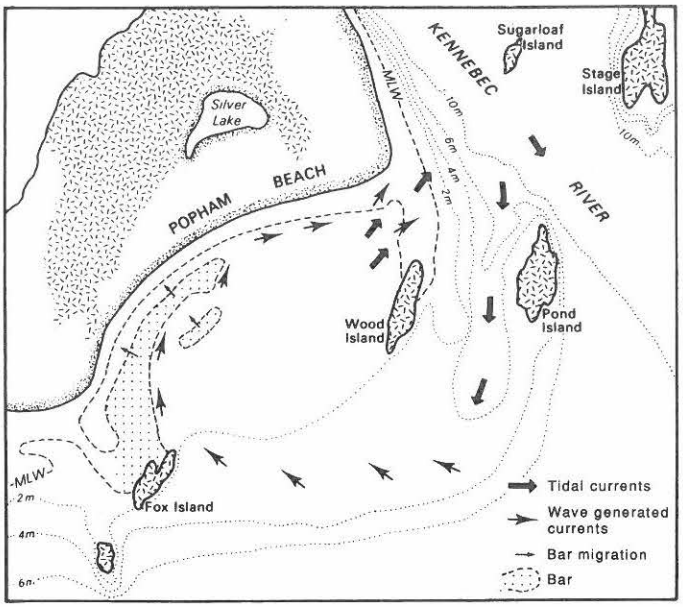

Figure 30. Sediment transport model for the mouth of the Kennebec River and Popham Beach (from FitzGerald and Fink, 1987).

the last 30 years confirms this. The stability of Scarborough Inlet argues that the majority of sediment in transport at the inlet is recirculated with little of it being deposited in long-term sinks. Even the Pine Point shoreline next to the south jetty has not built seaward appreciably during the last 10 years (Barker, 1988).

Sand transport patterns on the ebb-tidal delta are similar to those described for many other inlets (Bruun and Gerritsen, 1959; Dean and Walton, 1975; Hine, 1975; FitzGerald et al., 1976). Sand from the terminal lobe is moved onshore across the swash platform to the channel margin linear bars by wave-generated and flood-tidal currents. From there, it is dumped into the main channel where dominant ebb-tidal currents (Fig. 27) transport the sand to the terminal lobe, thus completing the sediment gyre. The presence of landward migrating swash bars at the Scarborough River Inlet indicates that sand is also exchanged between the ebb delta and adjacent beaches (Fig. 14) (Farrell, 1972; Barker, 1988).

The fact that Scarborough Inlet has a flood-tidal delta (Fig. 14), even though the inlet is dominated by ebb currents (Fig. 27), is not surprising. This same situation has been reported at several other locations including nearby Hampton, Merrimack, Parker, and Essex River Inlets (Boothroyd and Hubbard, 1975; Hayes 1975). These sand deposits will form if there is sufficient room. Sand is transported into the backbarrier primarily during storms when increased volumes of sand are delivered to the main channel and strong flood currents are produced by the storm surge.

Kennebec River Estuary

A large clockwise sediment gyre exists at the mouth of the Kennebec River which circulates sand along Popham Beach and in the offshore region (FitzGerald and Fink, 1987; Fig. 30). This movement of sand, which is driven by tidal currents near the mouth of the river and by wave-generated currents along the beach and on the offshore bar (Fig. 30), has been corroborated by bedform data, bar migrational trends, and beach surveys (FitzGerald and Fink, 1987). Along the southward-facing beach, sand is transported in a net easterly direction by dominant easterly longshore currents. Once in the vicinity of the Wood Island tombolo, flood-tidal currents augment the wave-generated currents, moving sand into the Kennebec River.

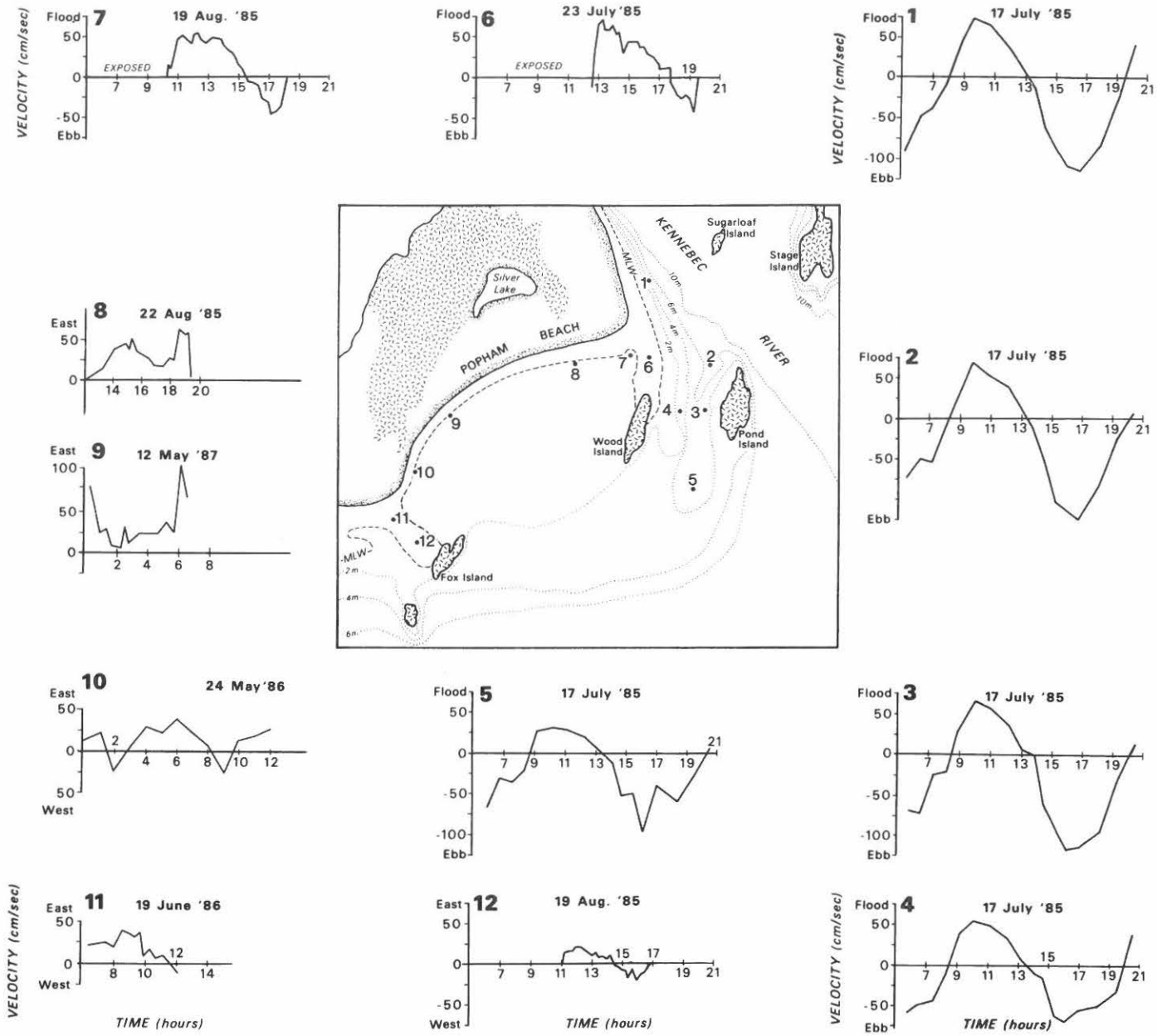

Figure 31. Representative current velocity curves for the Popham Beach region measured with a Marsh-McBirney electromagnetic current meter (data from FitzGerald and Fink, 1987).

Most of the sand delivered to the Kennebec River main channel appears to be transported into an ebb spillover lobe channel between Pond and Wood Islands (Fig. 31). A large, shallow terminal lobe (depth <2 m) has formed by the seaward movement of this sand. During easterly storms, waves breaking along the terminal lobe feed sand to a shallow bar that extends continuously to Fox Island, a distance of 1.5 km. Although not fully documented, aerial photography and some current data taken during storms indicate that sediment is moved in a westerly direction along this bar where it ultimately becomes incorporated into swash bars that form near Fox Island. These swash bars coalesce as they move onshore, forming a large bar complex that extends eastward along the beach up to 2 km and whose slipface is 2 to 3 m high (Fig. 32). The landward migration and attachment of the bar complex to the beach, having a volume of about 400,000 m^3 of sand, completes the clockwise sediment gyre.

Figure 32. Photographs of bar complexes migrating onshore along the southwest end of Popham Beach near Fox Island: (a) 1976 aerial photograph (taken by Fink) and (b) a ground photograph of the 2 to 3 m high bar slipface in 1982. These photographs were taken of two different bars. The bar present in 1976 had fully welded to the beach by 1978 and the bar present in 1982 had completely attached to the beach by 1984.

During the past 30 years, this process has been repeated every 6 to 10 years and has accounted for shoreline progradation and erosion amounting to over 100 m (FitzGerald and Fink, 1987).

SUMMARY

The occurrence of tidal inlets in Maine is closely related to the presence of large glacial deposits of sandy sediments. They are found in locations where sand supplies have been sufficient to form barriers: along the arcuate embayment shoreline of southern Maine and along the peninsula coast at the mouths of the Kennebec and Sheepscot Rivers. The morphology of Maine's inlets is dependent on their size which, in turn, is controlled by their tidal prism. Small and medium-sized inlets are anchored next to bedrock headlands and are bordered by spit platforms. They lack ebb-tidal deltas, but generally contain one or more well developed flood-tidal deltas. The backbarrier of these inlets consists primarily of high marsh and a central tidal creek with little intertidal sand flat or bay area. Large-sized inlets have similar morphology, but have better developed ebb deltas and more intertidal and bay area. In comparison to the tidal inlets, the Saco, Kennebec, and Sheepscot Estuaries have relatively less marsh and intertidal flat development but more open water area.

Tidal inlets in Maine have experienced relatively little historical migration and only Ogunquit River Inlet has actually changed location during the past 200 years. However, inlet processes have been and continue to be responsible for the greatest amount of shoreline change along the adjacent barriers. These processes, which include meanderings of the main ebb channel, spit accretion, and the exchange of sand between the ebb-tidal delta and landward beaches, are best exemplified at Morse River Inlet, Spurwink Inlet, and at the mouth of the Kennebec River, respectively.

Tidal hydraulics at Maine's inlets are strongly influenced by their size. At small to medium-sized inlets, the shallow nature of the main inlet channel causes a truncation of the lower portion of the ocean tide. This results in a prolonged ebb cycle, short flood cycle, and stronger mean and maximium flood than ebb current velocities. This condition produces net landward sediment transport into the backbarrier. At large inlets, the wider and deeper channel has less effect on the ocean tide and dominant ebb-tidal flow at these inlets can be explained in terms of inlet efficiency and filling characteristics of the backbarrier.

Due to Maine's indented coastline, sand transport patterns at its tidal inlets are characterized by discrete well-defined sand recirculation patterns rather than net movement out of the system. At small and medium-sized inlets these circulation cells include just the spit platform and inlet channel while at large inlets they normally encompass the entire ebb-tidal delta. At the mouth of the Kennebec River, tidal and wave processes produce a sediment gyre that involves a 1.5 km length of shoreline and is responsible for the formation and attachment of a bar complex to the beach that comprises over 400,000 m^3 of sand.

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