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Morphodynamics, stratigraphy, and sediment transport patterns of the Kennebec River estuary, Maine, USA

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Abstract

Sediment transport and circulation patterns within the lower Kennebec River estuary, Maine (~20 km) have been investigated over a two year and nine month period using fathometer profiles and side-scan sonograms in conjunction with flow measurements, fresh-water discharge data, and grain-size data. The geologic history of the estuary is inferred from high-resolution seismic profiles and bridge borings.

Subbottom data corroborate a five-stage evolutionary history that has been determined for other areas of the west-central Maine coast. Scattered deposits of glacial till (diamict) and stratified drift overly a Precambrian to Paleozoic metasedimentary bedrock basement. The glacio-marine blue clay of the Presumpscot Formation unconformably overlies the diamict and drift and drapes the basement, where till is absent. The clay surface is an erosional unconformity formed during the last sea-level lowstand. During subsequent sea-level rise, a relatively coarse-grained estuarine fill was deposited within a flood-dominated, relatively large paleo-Kennebec River estuary. As the rates of sea-level rise slowed, the system shifted to an ebb-dominated estuary in which the estuarine fill underwent reworking and downstream net transport.

Bathymetric data show a hierarchical arrangement of bedforms ranging in size and morphology from well-developed, ebb-oriented transverse bars to superimposed simple, straight-crested megaripples. The transverse bars were stable over the study period. The reworking and migration of the smaller forms are closely linked to seasonal variations in the relative contributions between tidal flow and fresh-water discharge. During the spring, large-magnitude discharge events augment ebb-tidal flows. The ebb-reinforced flows dominate the system and result in a net downstream transport of medium- to coarse-grained sand. Estuarine stratification plays an important role in sediment transport during non-spring months. From mid-summer to fall, salinity gradients enhance flood-tidal flows and result in minor quantities of upstream transport. In addition, bedrock bathymetric highs and abrupt changes in channel geometry may influence sediment transport within the estuary.

1. Introduction

The circulation and mixing of estuarine waters create highly variable physical regimes as reflected by the diversity of estuaries around the world. Due to the complex interactions among the dynamic physical processes operating within estuaries and at estuarine boundaries, our current understanding of estuarine process–response patterns is relatively limited. In addition, earlier studies of estuarine dynamics were primarily confined to Coastal Plain and micro-tidal environments (e.g., Meade, 1972; Ludwick, 1974). Later, an increase in the commercial

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development and utilization of estuaries stimulated studies which related estuarine circulation patterns to sedimentation patterns in mesotidal and macrotidal estuaries (e.g., Allen et al., 1980; Breusers and van Os, 1981; Gelfenbaum, 1983; Harris, 1988; Dalrymple et al., 1992).

For a variety of reasons, the Kennebec River estuary offers an excellent setting to study the hydrodynamics and sedimentation patterns within an elongated estuary. First, several episodes of Pleistocene glaciation have strongly influenced this high-latitude estuary. In particular, glacial scouring deepened and accentuated valleys along a metasedimentary and metavolcanic structural fold belt, and stripped away the existing sediment cover (Osberg et al., 1985). Subsequent fluvial erosion of unconsolidated glacial and periglacial deposits provided large volumes of sediment to the estuary. Consequently, this relatively unique rock-bound estuary, with a nonuniform channel geometry ranging from constricted and gorge-like to laterally expansive, possesses a highly variable temporal and spatial flow in which several morphologically distinct transverse bedform fields exist. Second, a high degree of tidal forcing and seasonally reinforced variations in the magnitude and frequency of fluvial discharge controls the estuary's hydraulic system. Last, uncertainty exists as to whether alluvial and estuarine sediment, glacio-deltaic sediment located ~8 km offshore of the present-day Kennebec River estuary mouth, or a combination of these potential sources may contribute sediment to some of Maine's indented coast beaches (FitzGerald and Fink, 1987; Belknap et al., 1989; Barnhardt et al., 1995).

Recently, Dalrymple et al. (1992) introduced a conceptual sedimentologic and morphologic classification of estuaries based on the relative importance of fluvial and marine processes. The degree of fluvial dominance determines the rate at which the estuary fills with sediment. Their models of the wave- and tide-dominated systems envision an inner estuarine zone controlled by riverine sediment transport, a central zone of low energy where river flow is countered by flood-tidal currents, and an outer zone dominated by waves and/or tides. The difference between the wave- vs tide-dominated estuarine end members is primarily reflected in the distribution of sand bodies at the mouth of the estuary. The wave-dominated estuary contains a barrier and tidal inlet complex at its mouth, whereas the tide-dominated estuary is fronted by intertidal sand bars that grade to mud flats and peripheral salt marshes in an up-estuary direction. Although the Kennebec River estuary contains elements of both the wave- and tide-dominated models, the components of the estuary do not fit neatly into Dalrymple et al.'s (1992) classification. The confining nature of the bedrock-cut channel precludes the formation of a central bay or the development of extensive intertidal sedimentary environments. Tidal flats and marshes occur only where small embayments border the main channel, particularly at the mouth. In addition, bedrock pinning points at the entrance to the estuary and an abundant sand supply, presumably from the upper estuary and perhaps offshore sources, have led to the formation of the bordering barrier system of Popham Beach (FitzGerald et al., 1989). Since the morphology and distribution of sedimentary facies in the Kennebec River estuary, as well as those in other New England estuaries, are largely controlled by the bedrock geology and structural grain of the region, a comprehensive estuarine model requires the inclusion of basement controls.

In this paper we examine the interaction between estuarine processes and sedimentation patterns within the Kennebec River estuary. Large volumes of sand- and gravel-sized sediments ubiquitously floor the lower portion of the estuary (approximately 20 km) and have been molded into a hierarchical organization (variety of superimposed scales) of bedforms. Periodic hopper dredging, needed to maintain federal navigation channels, attests to the highly dynamic substrate of the estuary (e.g., Science Applications Inc., 1984; Hubbard, 1986). By linking process data with response data over a variety of temporal and spatial scales, we show that seasonal fresh-water discharge events control net sediment transport patterns within the estuary.

2. Physical setting

The Kennebec River estuary receives fresh water from two major streams and their tributaries: the Kennebec and Androscoggin Rivers (Fig. 1). In terms of average annual discharge ($\overline{Q}$), the Kennebec River is the 19th largest river in the conterminous United States of rivers that drain directly

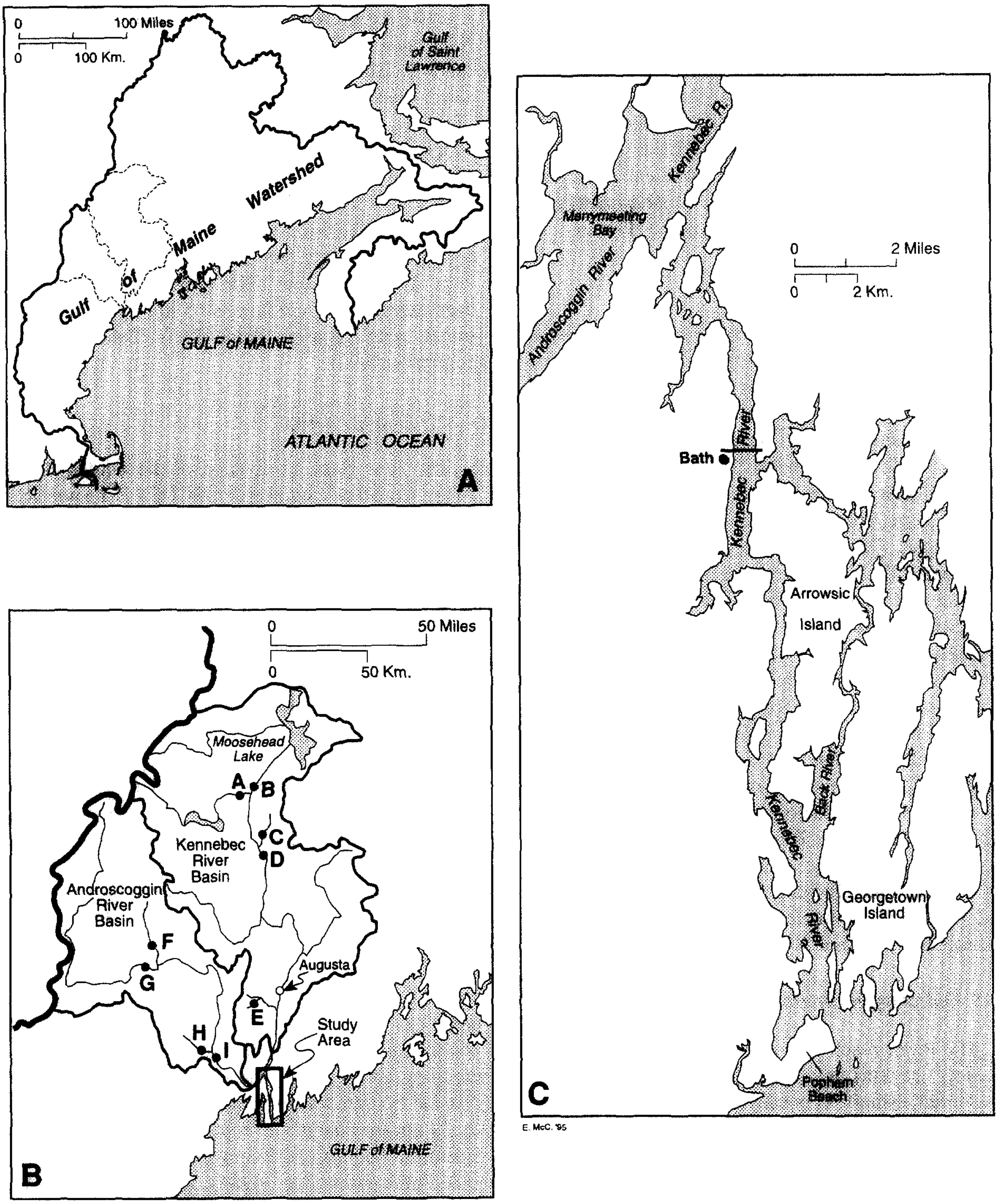

Fig. 1. Map of study area showing: (A) Gulf of Maine Watershed; (B) Kennebec–Androscoggin River watershed with United States Geologic Survey stream gaging stations A–I; and (C) lower Kennebec River estuary from the confluence of the Kennebec and Androscoggin Rivers at Merrymeeting Bay to the mouth near Popham Beach.

into the ocean ($\overline{Q} = 280.3$ m^3 s^{-1}; Nace, 1970). It has a drainage area of approximately 14,775 km^2 and its 225 km length begins at Moosehead Lake in west-central Maine. The Androscoggin River is 213 km long, has a drainage area of approximately 8500 km^2 and heads in Umbagog Lake in northeast New Hampshire. The Androscoggin and Kennebec Rivers join at Merrymeeting Bay approximately 20 km north of the Kennebec River mouth (Fig. 1).

2.1. Sediment supply

The modern-day (and throughout post-glacial time) coarse-grained sediment contribution to the Kennebec River estuary comes primarily from fluvial erosion of unconsolidated glacial ice-contact and periglacial deposits. In particular, outwash deposits within the Kennebec River drainage system, such as those of the Embden Formation, have furnished large volumes of sand to the river (Borns and Hagar, 1965). Likewise, along the Androscoggin River, the channel has scoured into extensive glacial sand deposits (Thompson and Borns, 1985) such as the Brunswick sand plain (T. Weddle, pers. commun., 1995). The coarse-grained nature of these two river systems is exhibited at Merrymeeting Bay where extensive sand shoals are exposed at low water. These deposits are probably mobilized during high discharge, high flow velocity events, and the eroded sands are transported to the lower Kennebec River. Erosion of glacio-marine silts and clays (Presumpscot Formation; Bloom, 1963) along both river courses supply fine-grained sediment to the system. In addition, an offshore paleodelta and littoral areas have been postulated to supply sediment to the estuary (Belknap and Kelley, 1987).

2.2. Hydrology/hydrography

The lower reach of the Kennebec River (approximately 20 km) is a partially-mixed to stratified mesotidal estuary (Fenster et al., 1987). The degree of mixing within the estuary varies along the length of the river depending on the magnitude of fresh-water discharge and the tidal phase. Seasonal variations in river discharge exist with minimum flow occurring in late summer and mid-winter and maximum flow conditions occurring in early winter and late spring. Tides are semidiurnal and the mean tidal range at the mouth of the Kennebec River estuary is 2.6 m, increasing to 3.5 m during spring tides. These conditions produce a mean tidal prism of $1.01 \cdot 10^8$ m^3 which is 16 times greater than the average fresh-water discharge over the same period (FitzGerald et al., 1989). Consequently, the lower portion of the Kennebec River estuary experiences reversing tidal flow, except during extremely large spring freshets (e.g., the 1987 April flood) and other large discharge events. The limit of the tidal influence during low river flow is 65 km upstream of its mouth near Augusta, Maine (D.W. Caldwell, pers. commun., 1995).

2.3. Dredging history

The New England division of the United States Army Corps of Engineers (USCOE) is responsible for maintenance dredging operations of the Kennebec River federal navigation channel. Two regions of the Kennebec River estuary require dredging for deep draft vessels entering and leaving the Bath Iron Works Shipyard (Normandeau Associates, 1994). An 8.2 m deep (at mean low water) channel is maintained in the immediate vicinity of Bath Iron Works (~17 km north of mouth). Near the mouth of the Kennebec River estuary, adjacent to Popham Beach, a 1.1 km long channel is also dredged to a depth of 8.2 m. Dredge material from the northern site is deposited within the estuary approximately 2.8 km south of the dredge site. The dredge spoil for the southern site is dumped approximately 3.3 km offshore of the Kennebec River estuary mouth.

The northern dredge site requires dredging every one to four years and on the average approximately 42,700 m^3 is removed. At the mouth of the river, the channel is dredged about every 12 years and an average of 47,900 m^3 of sand is removed. A 4-month post-dredging study on the effectiveness of the 1981 dredging operation (northern site) indicated that the channel bottom was extremely dynamic and subject to significant changes in topography most likely as a result of bedform development and migration (Science Applications Inc., 1984). An additional study, which monitored the northern disposal site, showed that, over the 11-month survey period, sand moved downstream a net distance of 300 m (Hubbard, 1986).

Table 1
Bedform orientations for subreaches within the Kennebec River estuary for the 12 survey dates relative to tidal range, tidal stage during survey, and fresh water discharge (Q_f)

Date	Survey type, survey length (km)	Bedform orientation reach					Tidal range (predicted) (m)	Tidal stage (m)	Q_f (cm)
		I	II	III	IV	V			
30 April 1986	F, 20	E	E	V	V	E	1.86	1/2 ebb	328
27 June 1986	F, SS, 20	V	F	F	V	V	2.04	1/2 flood	144
4 July 1986	F, 20	V	V	F	V	V	1.49	1/2 flood	163
2 August 1986	SS, 20	F	V	V	F	V	1.00	1/2 flood	217
1 October 1986	F, 3.2	F					1.77	Profile #1; ebb/slack	150
		F					1.68	Profile #2; 3/4 flood	
		F					1.62	Profile #3; 1/2 ebb	
		F					1.62	Profile #4; 3/4 ebb/slack	
4 April 1987	F, 20	E	E	E	E	E	1.00	1/2 ebb	1211
13 August 1987	F, 20	F	V	F	E	E	2.50	>3/4 flood	130
14 August 1987	F, 20	F	V	F	E	E	2.26	3/4 ebb	128
2 October 1987	F, 20	F	V	F	V	V	1.71	<1/4 flood	78
30 October 1987	F, 3.2	F					1.31	Profile #1; 1/2 ebb	211
		F					1.31	Profile #2; >3/4 ebb	
		F					1.62	Profile #3; 2/3 flood	
13 April 1988	F, 20	E	E	E	E	E	2.47	1/2 ebb	365
30 January 1989	F, S, 24	F	V	F	F	F	1.00	1/2 flood	153

Note that bedforms do not change their orientation over the course of one or sequential tidal cycles. Rather, the bedforms are ebb-oriented during the spring months, as the fresh-water discharge increases by a factor of two to ten, thereby supplanting the flood-tidal currents as the dominant transport mechanism. Bedforms remain flood-oriented during low Q_f conditions of summer and early fall. Thus, the greatest amount of downstream movement of all large-scale bedforms (sand waves and transverse bars) in the Kennebec River estuary occurs during high-magnitude spring freshets and other large flood events. F = fathometer; SS = side-scan sonar; S = seismic survey. Reaches: I = Bath to Doubling Point; II = Doubling Point to Bluff Head; III = Bluff Head to Squirrel Point; IV = Spuirrel Point to Perkins Island; V = Perkins Island to Fort Popham (see Figs. 2 and 4 for locations) (E = ebb; F = flood; V = variable).

3. Methods/data base

Over a two year and nine month period (April, 1986 to January, 1989), we gathered a seasonal set of remotely sensed data consisting of 12 surveys along all or a portion of the 20 km lower estuary (Table 1). These data were supplemented with sediment samples, data from bridge borings, and hydrographic surveys.

3.1. Bathymetric data

Narrow-beam fathometer and side-scan sonar surveys were used to obtain information on: (1) the spatial scales of the bedforms within the channel; (2) the temporal dynamics of bedforms comprising the estuary bed (e.g., over a tidal cycle and seasonal); and (3) the channel geometry (i.e., depth and cross-sectional profiles). Longitudinal bathymetric profiles were collected during spring, summer, and fall months from 1986 to 1989. Except for a January 1989 fathometer and seismic survey, winter data were difficult to obtain due to extreme conditions (e.g., floating ice blocks).

Two sets of longitudinal surveys were conducted. The first set consisted of eight fathometer surveys and two side-scan sonar surveys (using a Raytheon DE-719B Precision Depth Recorder, and EPC and Klein systems, respectively). Each survey covered a distance of approximately 20 km from the Carlton Bridge (Route 1) at Bath, Maine to the estuary mouth near Fort Popham (Fig. 2A; Table 1). The second set of surveys consisted of repetitive fathometer profiles taken over the course of a tidal cycle from the bridge at Bath to the Doubling Point Lighthouse (covering a distance of 3.2 km, Fig. 3A). Ship tracklines were replicated as closely as possible using buoys and other markers in order to determine bedform orientations and migration patterns. Consequently, we could use bedform orientation as an indicator of sed-

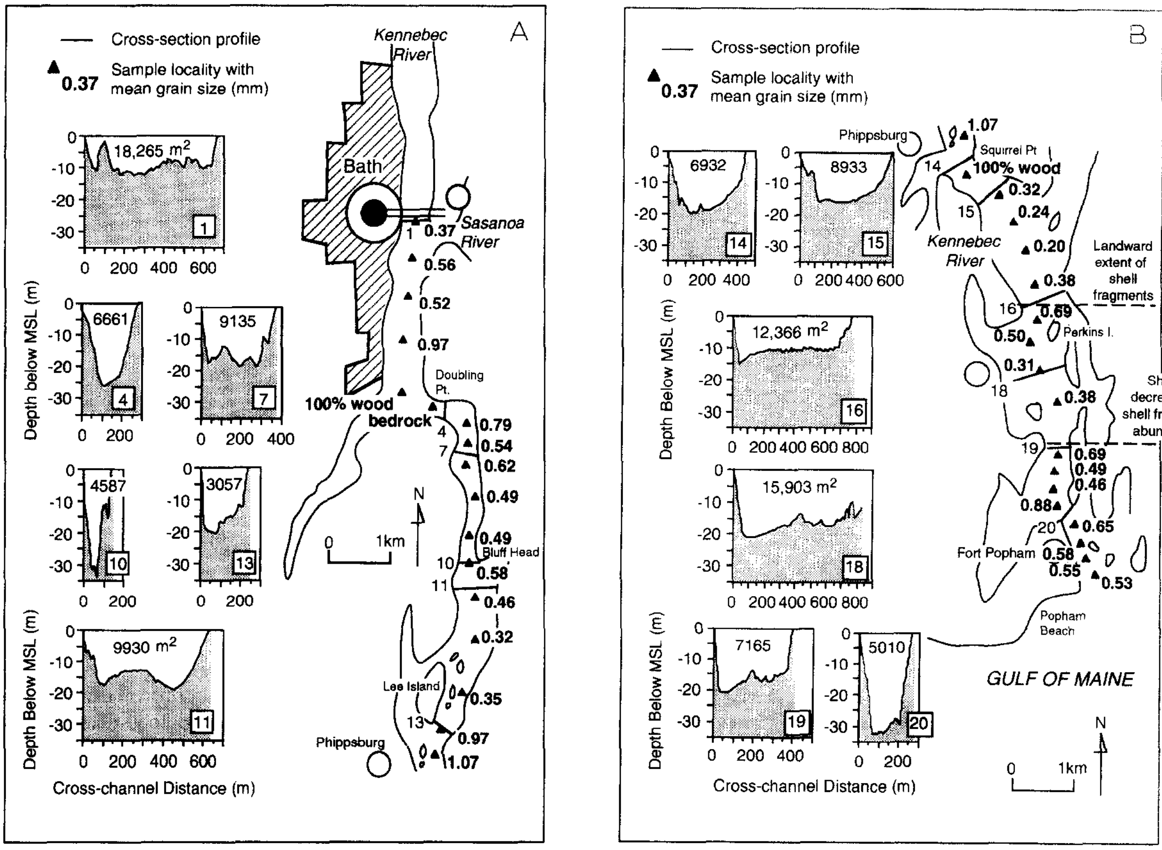

Fig. 2. Grain-size data and cross-sectional areas for (A) the northern and (B) the southern part of the Kennebec River estuary. Note the predominance of medium- to coarse-grained sands and the highly variable channel geometry.

iment transport directions. Finally, cross-sectional areas were determined at 20 locations along the estuary using fathometer data (Fig. 2). All bathymetric data were digitized and corrected for differences in boat speed and tidal height.

3.2. Sediment-size data

A Van Veen grab sampler was used to collect 31 sediment samples along the length of the estuary at regularly spaced locations (Fig. 2). Samples were collected during slack water as close to the channel thalweg as possible. The samples were processed and analyzed in the laboratory according to standard, referenced procedures using an Ottoman-type sample splitter and a mechanical shaker (e.g., Pettijohn, 1957; Folk and Ward, 1957). During washing of the sediment on a 4-ϕ sieve we found that none of the samples contained enough silt or clay for analysis. Descriptive statistical measures of grain-size distributions were determined using the Folk and Ward (1957) method.

3.3. Subbottom data

High-resolution subbottom acoustic (seismic-reflection) records were obtained along the 24 km long longitudinal axis of the estuary from Merrymeeting Bay to the mouth of the Kennebec. A Geopulse seismic system was used with a 1.5 kHz band pass providing a vertical resolution of 0.5 m. Depths below mean sea level to major acoustic reflectors and thicknesses of inferred sedimentary units were determined using a compressional-wave velocity of 1.5 km s^{-1} for water and Holocene sediments. Ship speed (Bath Iron Works Tug *Kennebec*) aver-

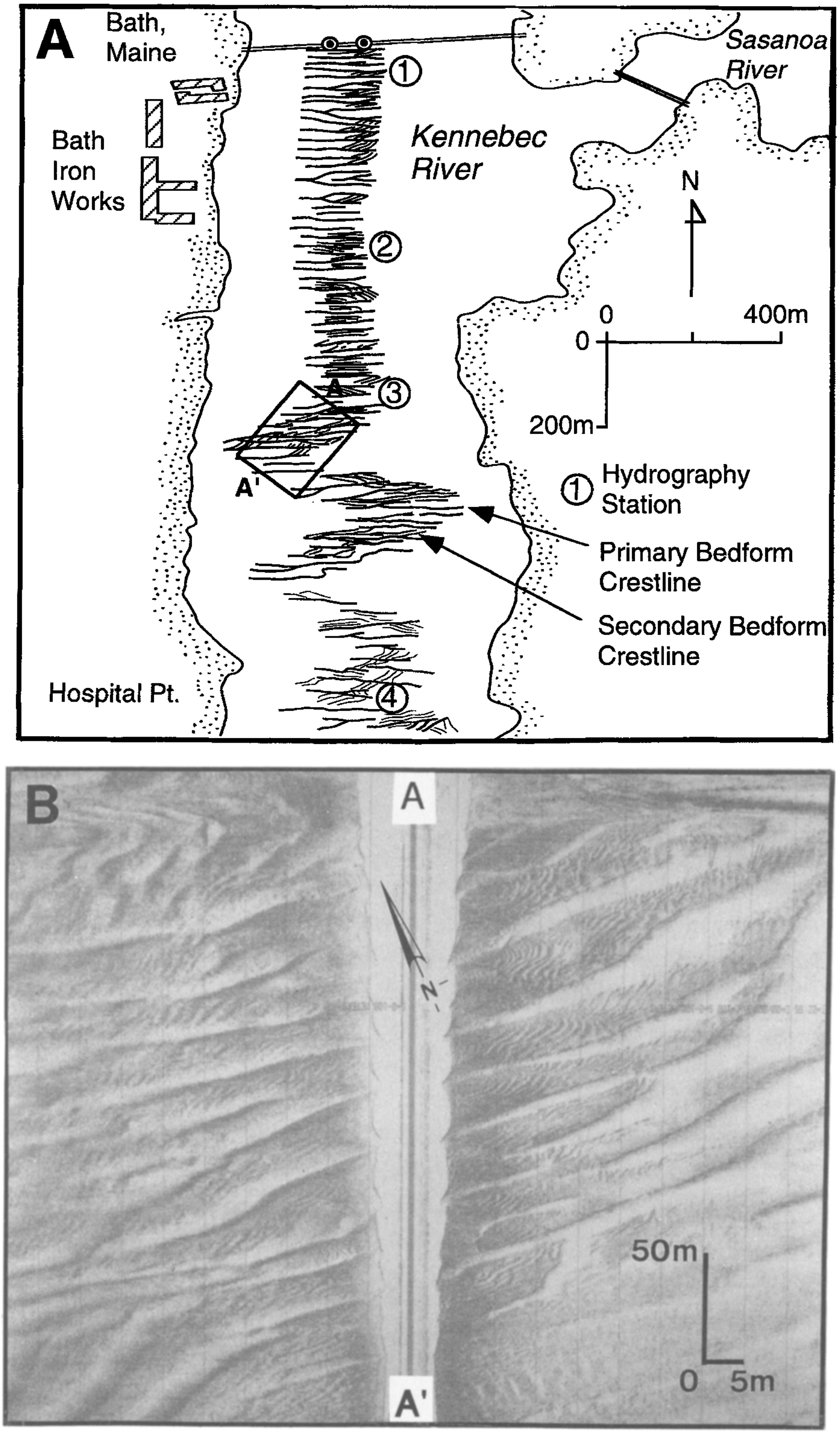

Fig. 3. (A) Primary and secondary bedform crestlines obtained from 27 June 1986 EPC side-scan sonar cruise. (B) Partial sonogram showing megaripples superimposed on sand waves (see inset in (A) for location). Note that the floor from Bath to Doubling Point contains sinuous, bifurcating, and straight-crested sand waves with oblique megaripples on their stoss sides. Numbered circles are locations of hydrography stations used in this study.

aged approximately 5 knots (9 km/h). Ship position was determined by line of sight. The narrowness of the channel and horizontal control obtained from earlier surveys produced a positioning accuracy comparable to that of Loran C. Stratigraphic control was available in the Bath region from a subsurface investigation for the Carlton bridge (core borings) by the Maine Department of Transportation.

3.4. Hydrographic data

3.4.1. Estuarine measurements

Hydrographic measurements were taken approximately one year apart (1 October 1986 and 30 October 1987) at four stations in the lower Kennebec River estuary (16.2 km to 19.6 km upstream from the mouth; Fig. 3A). Salinity, water temperature, and current velocity measurements were obtained repeatedly at each station over a 12-h period (one half a tidal cycle) with a VSI salinometer and a Marsh–McBirney current meter. Readings were made in 2 m intervals from the bottom to the surface. Tidal height was measured concurrently with the hydrography data from a tide gauge located on a shore piling.

3.4.2. Stream analyses

In order to quantify fresh-water discharge in the Kennebec River–Androscoggin River drainage basin, stream flow data obtained from nine United States Geological Survey stream gauging stations were analyzed using a number of standard surface water analytical techniques. The analyses yielding the greatest amount of information included: (1) regression analysis of long-term mean annual discharge vs drainage area; (2) flood-frequency analysis and homogeneity test; and (3) seasonality analysis.

The relationship between long-term mean annual fresh-water discharge and drainage area for various stream gauging stations within the Kennebec–Androscoggin watershed was plotted and expressed as a power function equation of the general form:

$$Q_{\mathrm{ma}} = cA_{\mathrm{d}}^{f} \quad (1)$$

where Q_{ma} is the average annual discharge, c is the value of the ordinate intercept when $x = 0$, A_{d} is the drainage area, and f is the slope of a least squares regression line. This analysis permits the estimation of average annual discharge at ungauged sites given the drainage area. The average annual discharge at the Kennebec River estuary mouth was determined based on a drainage basin area of 14,775 km^2 (Nace, 1970). In addition, discharge values were estimated at different regions within the estuary, using a similar analysis for daily flows during each of the survey and hydrography dates.

For the seasonality analysis, an annual probability hydrograph of a water year was constructed from mean monthly discharge values for one location on the Kennebec River watershed (gauge D at Bingham, Maine; drainage area = 7076 km^2; Fig. 1B) and one on the Androscoggin River watershed (gauge I near Auburn, Maine; drainage area = 8473 km^2; Fig. 1B). Due to the similarity of the results, cumulative frequency distributions of mean monthly discharge were plotted for one station only (gauge I).

4. Results

4.1. Channel morphology

Water depths in the thalweg range from 6 m in regions of active sediment transport to 30 m where the estuary is constricted by bedrock (e.g., near the estuary mouth). Cross-sectional areas of the channel vary from 18,265 m^2 near Bath, Maine, to 3057 m^2 at the southern end of Lee Island (Fig. 2). From Merrymeeting Bay, the Kennebec River estuary channel is highly variable displaying both narrow, gorge-like and wide, shallow geometries (Figs. 1, 2). Hydraulic regimes and sedimentation patterns are highly variable due to this broad range of channel geometries.

4.2. Bedform dynamics

A hierarchical suite of sedimentary features composed of medium- to coarse-grained sand are present within the lower kennebec river estuary (Figs. 3, 4). These features range from simple, straight-crested megaripples to complex systems composed of 'bars' with superimposed sand waves and megaripples. (According to the classification scheme of Ashley (1990), 'bars' with superimposed sand waves and megaripples are two-dimensional, very large, compound subaqueous dunes; sand waves are two-dimensional, large to very large dunes, and megaripples are two- and three-dimensional, small to

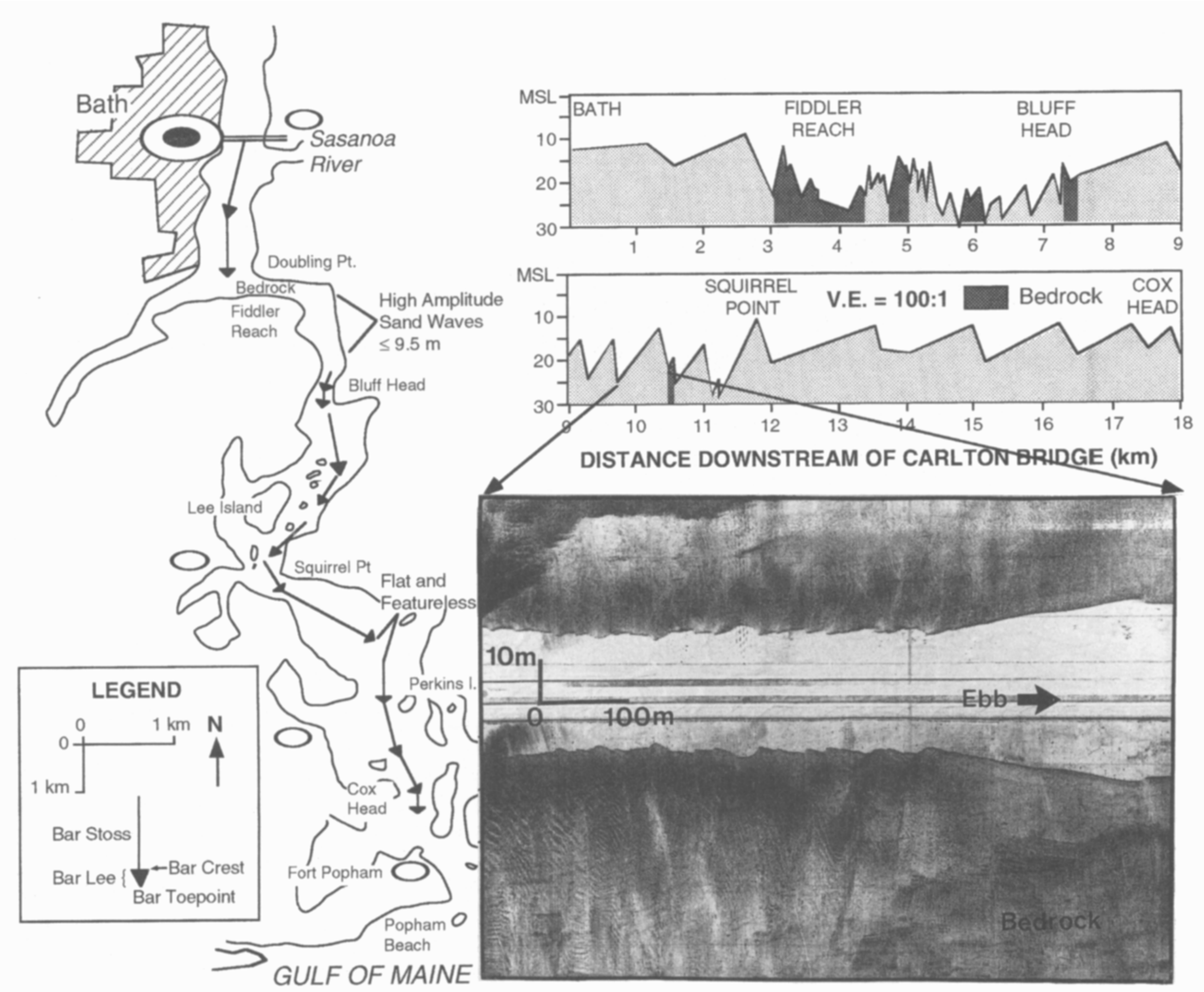

Fig. 4. Location and longitudinal bathymetry of ebb-oriented transverse bars within the Kennebec River estuary. Inset shows swath of Klein side-scan sonogram with three orders of hierarchical sedimentary features: transverse bar with superimposed sand waves and megaripples (sonogram obtained north of Squirrel Point).

medium, subaqueous dunes.) The largest features are ebb-oriented and transverse to the predominant flow directions with maximum heights of 10 m and wavelengths ranging from 0.4 to 1.2 km. Secondary forms consist of sand waves with maximum heights and wavelengths of 6.5 m and 50 m, respectively. Third-order bedforms are composed of megaripples with heights of 0.2 m to 0.6 m and wavelengths of 2.0 m to 3.0 m. The superimposed forms occur on the stoss-side and in the troughs of the larger forms and are commonly oriented obliquely to the larger bedforms. Occasionally, active sand waves with well-defined slip-faces were observed on the lee slopes of the bars.

The bathymetric data show that a series of large-scale transverse bars extends the length of the study area (Fig. 4). These bars display typical gentle-sloped stoss-sides (1° to 3°) and slightly steeper lee-faces (4° to 6°). Although smaller-scale bedforms can migrate over both stoss- and lee-sides, the forms maintained their ebb-orientation over the study period. Due to limitations involved in the data collection procedure (e.g., navigational inaccuracy), rates of bedform movement could not be computed.

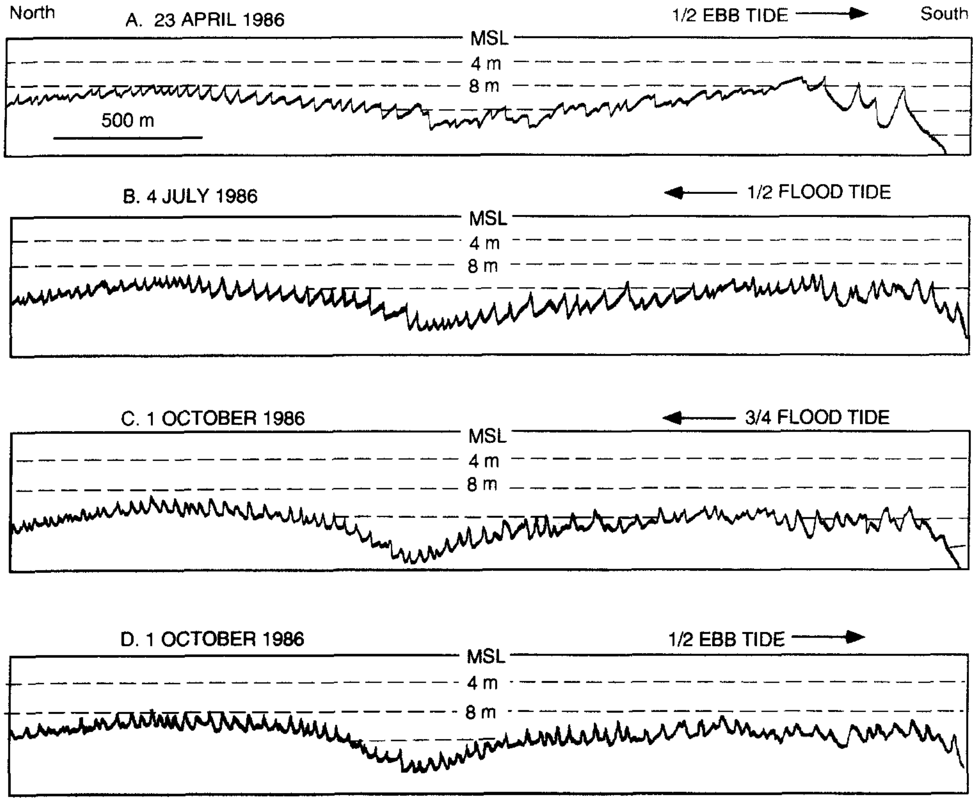

Fig. 5. Fathometer profiles of Bath to Doubling Point from 23 April 1986 (A) and 4 July 1986 (B). North is to the left, south to the right. All bedforms in 23 April survey ebb-oriented and bedforms in July range from ebb- to flood-oriented although residing in a flood tide. Repetitive fathometer profiles of Bath to Doubling Point for 1 October 1986 (C and D). Although profiles were obtained from two different portions of the tidal cycle, sand wave orientations did not change.

In general, the bar crests are located in the vicinity of abrupt changes in channel morphology, for example, in regions where the ebb-flow is greatly reduced upon entering a wide channel section.

Table 1 lists observations made regarding smaller-scale bedform dynamics for the Bath to Doubling Point reach (see Figs. 2A, 3A for locations). Sand wave morphologic and orientation changes occur over much shorter time scales than the bars. However, the bathymetric data indicate that sand wave orientation does not correlate with diurnal or fortnightly tidal cycles (Fig. 5; Table 1). Rather, the sand waves maintained a seaward (ebb-) orientation throughout the spring and early summer months (Figs. 5A, 5B; Table 1). A change from ebb- to flood-orientations in the middle to late summer, and back to ebb-orientations in the late winter to early spring, suggests that seasonal processes control bedform migration patterns (Table 1). We provide evidence below which indicates that the primary process responsible for these patterns is the seasonal variation in fresh-water discharge. During the spring months, fresh-water discharge increases by a factor of two to more than ten (during large-magnitude events), thereby supplanting the flood-tidal currents as the dominant transport mechanism (see Table 1 and Sect. 4.5.2).

Bathymetric data obtained from repetitive surveys over the course of a half tidal cycle (Figs. 5C, 5D; 1 October, 1986), and from surveys on consecutive days (Table 1; 13–14 August, 1987) further corroborate the finding that the bedforms do not change their orientation over the course of one or sequential tidal cycles. Occasionally, over a tidal cycle, the upper third (shallowest parts) of some of the smaller forms appeared to steepen and become symmetrical.

It should be noted that only one day after the federal navigation channel was dredged to the 9.8 m isobath below mean low water (1 October, 1986), the bedform field had become reestablished (Fig. 5C, D). Thus, the repetitive survey data indicate that, although the bedform crests can be reworked, bedform asymmetry changes and net bedform movement are unaffected by diurnal flow regime changes.

Table 1 gives bedform orientations for subreaches within the estuary for each available survey (see Fig. 2 for locations). These results show that bedform orientation is highly variable along the longitudinal axis of the estuary and can vary between or within subreaches. For example, the upper reach can possess flood-oriented sand waves (i.e., Bath to Doubling Point), while the lower reaches can exhibit ebb-oriented sand waves (e.g., Squirrel Point to Fort Popham) (cf., 13–14 August, 1987; Table 1). Consequently, the general patterns discussed above can vary as a function of channel geometry and sheltering by bathymetric highs (e.g., bars and bedrock outcrops).

4.3. Sediment-size analysis

Medium- to coarse-grained, moderately-well to well sorted sand dominates the Kennebec River estuary, although the sample obtained from a channel gorge adjacent to Fort Popham contained 99% shell fragments and other samples from the lower estuary contained lesser amounts of shell material (Fig. 2). The mean grain size of the Kennebec River estuary is 0.48 mm (1.05 ϕ). In some of the bedrock depressions samples contained 100% wood fragments (e.g., near Squirrel Point and in the narrow reach between Doubling Point and Fiddler Reach Point). We duplicated the grain-size analysis for a sample obtained south of Squirrel Point to check for consistency in our methods (and results) and found that the mean size for each subsample was 1.63 and 1.67, with corresponding sorting values of 0.50 and 0.49, respectively.

As expected, a general relationship exists between the width of the channel and the sediment characteristics (Fig. 2). In wider reaches, grain size decreases and sorting increases. However, little correlation was found between mean grain size and sorting ($y = -0.1x + 0.7$; $r^2 = 0.14$).

4.4. Subbottom geology

The continuity, amplitude, frequency and spacing of subbottom reflections provide a means of defining seismic facies units within the Kennebec River estuary (Mitchum et al., 1977; Sangree and Widmier, 1979). Correspondingly, we interpret vertical facies changes from distinct differences in internal seismic character or a marked reflection horizon representing an erosional unconformity or correlative conformity.

Stratigraphically, the lowest prominent horizon recognized in the Kennebec River estuary channel is a strong reflector of highly irregular relief (Fig. 6). This irregular surface extends the length of the estuary. The seismic records of the basement unit are typically reflection-free. Bridge boring data (Carlton Bridge) indicate that the basement unit in the vicinity of Bath, Maine is a weathered biotite and muscovite schist (State Highway Commission, 1957).

In isolated regions, a unit characterized by chaotic reflectors and a strong upper boundary reflector overlies the bedrock basement. We assume that this seismic facies correlates with the till (basal and moraine) and stratified drift found in nearby Sheepscot Bay (Belknap et al., 1986, 1989).

Where the inferred till unit is absent, a seismic unit characterized by acoustical transparency overlies the bedrock basement (Fig. 6). Where present, acoustically laminated reflectors within this unit mimic the underlying topography (Fig. 6). The thickness of this unit ranges from 0 to 23 m. Bridge boring data indicate this unit consists primarily of 'blue clay' of the Pleistocene-aged Presumpscot Formation (Fig. 7; Waddell, 1925; Bloom, 1963). The clay unit is punctuated by a ubiquitous, relatively flat-lying, strong, continuous seismic reflector at a depth of ~ -20 to -26 m. In places, the internal acoustic laminations of the clay unit are truncated by the strong reflector. Belknap et al. (1986) found evidence of channelization in, and scour lag on, the blue clay in the adjacent Sheepscot Estuary using a transverse survey. Although we could not identify similar features in our longitudinal survey of the Kennebec River estuary, the bridge boring data suggest the presence a subbottom channel (Fig. 7).

A seismic facies, having a variable thickness (0–15 m) and characterized by discontinuous, discordant internal reflections, unconformably overlies the blue

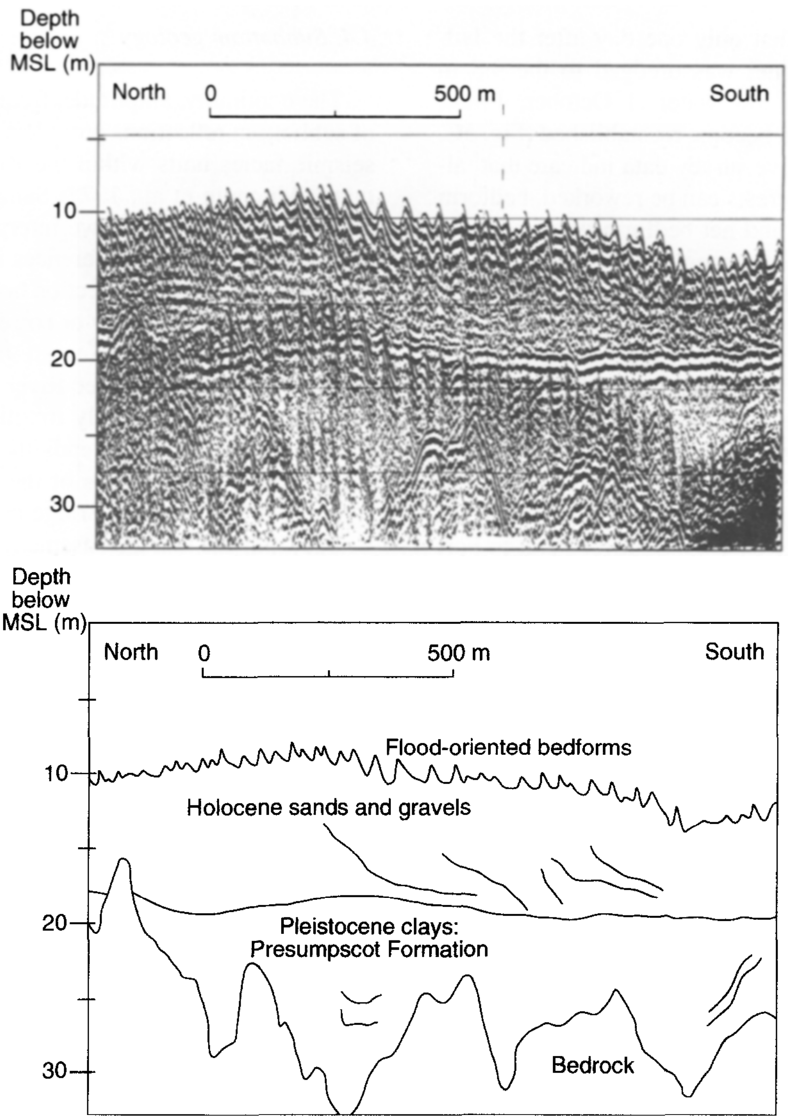

Fig. 6. Longitudinal seismic trace and interpretation obtained near Bath, Maine. The tripartite seismic stratigraphy consists of a reflection-free basal unit interpreted as the basement Paleozoic bedrock unit, overlain by an acoustically transparent (with occasional acoustically laminated reflectors) of glacio-marine clays which, in turn, is overlain by a chaotic seismic reflection pattern interpreted as reworked Holocene sands. Note the relatively flat-lying, strong (high amplitude and frequency), continuous seismic reflector at a depth of ~ -19 m separating the Pleistocene clays from the Holocene sands.

clay (Fig. 6). This internal reflection configuration suggests that deposition occurred in a moderate- to high-energy environment and/or disruption of beds occurred after deposition (Sangree and Widmier, 1979). In most places along the channel's longitudinal axis, the upper surface of this unit is highly irregular due to the presence of bedforms. However, distinguishable seaward-dipping tangential reflectors (dip angles <6°) were observed within some of the bar forms and below the region of active sand waves (Fig. 6). In addition, near the mouth of the estuary, a prominent (high amplitude and frequency), relatively flat-lying, continuous reflector emanates from the trough of a transverse bar, and extends upstream approximately 1.5 km of the bar's toepoint.

Our grab samples and the bridge boring data indi-

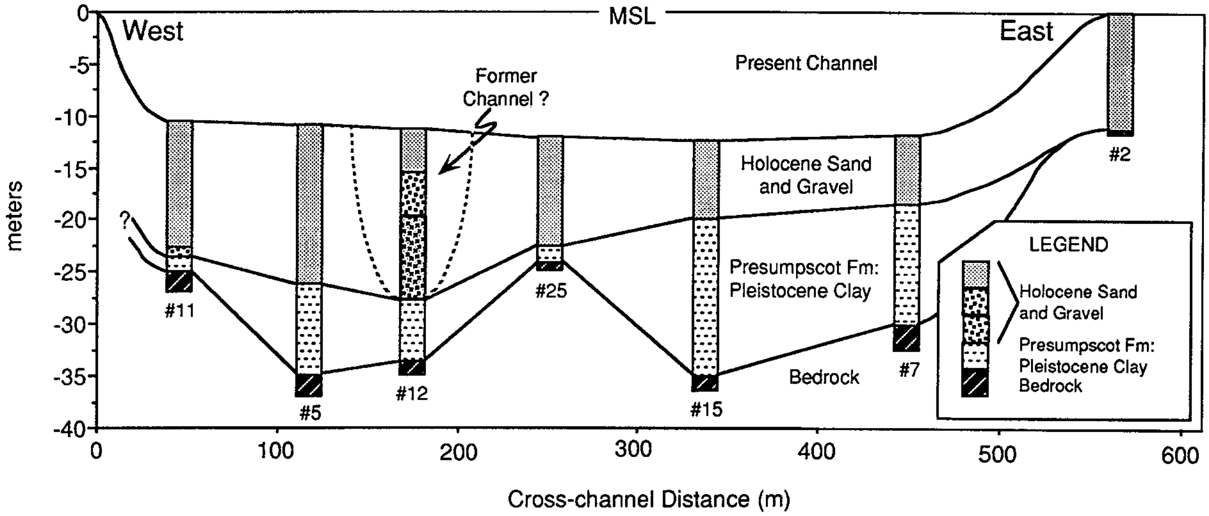

Fig. 7. Boring data from the construction of Carlton Bridge at Bath, Maine. The borings were obtained across the channel (east–west) just south of the present-day bridge location corresponding to cross-section 1 in Fig. 2A. The average cross-channel thickness of the Holocene deposit is approximately 11.7 m.

cate that the upper seismic unit consists of medium- to coarse-grained sands and gravels (Figs. 2, 7). We interpret this unit, which stratigraphically and unconformably caps the glacial sediment (and overlies the basement where glacial sediment is eroded), as coarse-grained estuarine sediment deposited during the middle phases of the Holocene transgression (see Sect. 5). In regions where the channel is particularly narrow (i.e., 250 m in Fiddler Reach; see Fig. 4 for location), all semi-consolidated and unconsolidated sedimentary units are absent and the bedrock basement forms the channel bottom.

4.5. Hydraulics (hydrodynamics)

4.5.1. Tidal flow

During both hydrographic surveys the currents at the four stations between the Bath bridge and Doubling Point generally exhibited stronger maximum (depth-averaged) ebb than flood velocities (Fig. 8A, 8C). For example, at Station 2, maximum ebb currents were 1.19 m s^{-1} and 0.95 m s^{-1} for 1 October 1986 and 30 October 1987, respectively, and maximum flood velocities were 0.82 m s^{-1} and 0.73 m s^{-1} during the same period (Fig. 8A, 8C). Whereas the depth-averaged values indicate a dominance of ebb-directed flow in this part of the estuary, measurements taken 2 m off the bottom suggest a different pattern. For example, on 1 October, near-bottom flood velocities exceeded those of the ebb at all stations except for Station 4, and the near-bottom flood currents at Stations 1 and 3 were more than twice as strong as the ebb currents (Fig. 8B). Maximum velocities for the ebb and flood tide occurred approximately two hours prior to slack water.

4.5.2. Stream flow

Fig. 9 shows the log–log relationship between mean annual discharge and drainage area for the Kennebec River and the Androscoggin River. The exponents of the power function equation are 0.97 and 0.98, respectively indicating: (1) the larger streams in the watershed have approximately the same amount of discharge per square area as smaller streams; and (2) the catchment area is hydraulically homogeneous.

As mentioned, one advantage of Q_{ma} vs A_d plots is that the average annual discharge at ungauged sites can be estimated. Using the value of 14,775 km^2 for the Kennebec River drainage basin area (Nace, 1970), the average annual discharge at the Kennebec River mouth is 261 m^3 s^{-1}. The y-intercept values of Fig. 9 (0.0220 for the Kennebec River and 0.0223 for the Androscoggin River) are the discharge values (in m^3 s^{-1}) per km^2. The values of the Kennebec River and Androscoggin River are similar to other New England streams which are subject

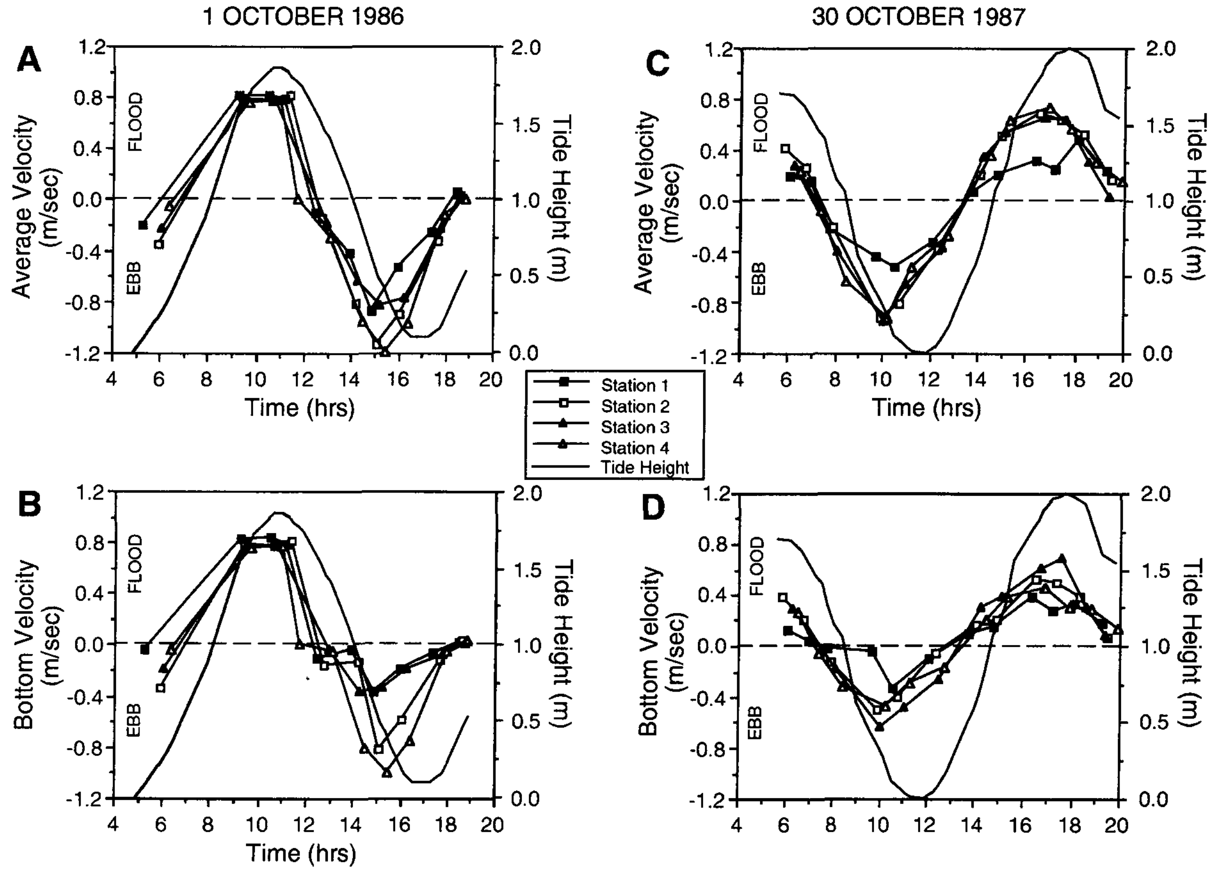

Fig. 8. Current velocity and tidal height measurements over a tidal cycle for 1 October 1986 (A and B): (A) depth-averaged velocity showing stronger ebb-current maximum velocities than flood-current velocities; and (B) bottom current velocities showing stronger flood currents at two stations (1 and 3). Current velocity and tidal height measurements over a tidal cycle for 30 October 1987 (C and D): (C) depth-averaged velocity showing stronger ebb-current maximum velocities than flood-current velocities; and (D) bottom current velocities showing slightly larger flood velocities than ebb. See Fig. 3A for location of hydrographic survey stations.

to large-magnitude freshets. The similarity between the Kennebec River function and the Androscoggin River function is due to their proximity and similar climatic conditions.

There are two main causes of flooding in New England: (1) intense rainfall associated with extra-tropical storms and hurricanes; and (2) freshets produced by rainfall coupled with melting of winter snow cover. The latter phenomenon is the most common cause of flooding in New England. One example of large-magnitude flooding occurred in April, 1987 during the course of this study (Fontaine, 1987; Stumpf and Goldschmidt, 1992). Record rainstorms in conjunction with high temperatures melted record snow pack accumulations in a relatively short time period (Fig. 10A). The annual flood analysis showed that the April 1987 freshet was the largest flood event in recorded history at most of the gauging stations (Fontaine, 1987; Stumpf and Goldschmidt, 1992). For example, the discharge during the peak of the early April flood was 6230 m^3 s^{-1} (2 April) on the lower Kennebec River, 194% greater than the previously recorded largest annual flood of 3200 m^3 s^{-1} (1 June 1984).

As expected, the seasonality analysis shows that high water discharge periods occurred in the spring months and early winter (Fig. 10B). Spring high streamflows are directly proportional to freshet events. Early winter high events are related to increased stormflow (decreased infiltration) during the

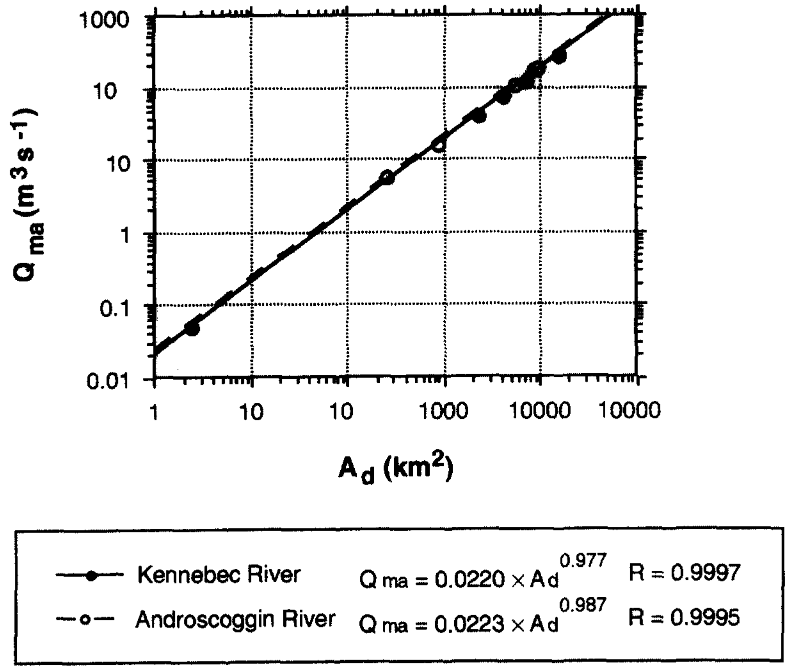

Fig. 9. Long-term mean annual discharge (Q_{ma}) as a function of drainage area (A_d) for various stream gauges in the Kennebec–Androscoggin watershed. Note the similarity between the Kennebec and the Androscoggin Rivers in discharge versus drainage area.

first frost season. Low-water seasons occur prior to high-water seasons. Winter low discharge events in February are related to storage of water as snowpack from January to March. Summer lows are due to decreased precipitation (lowering of the water table) and increased evapotranspiration.

4.5.3. *Temperature and salinity variations*

Significant differences existed between the salinity data for 1 October 1986 and 30 October 1987 (Fig. 11; see Fig. 3A for station locations). Salinity values for October 1986 varied with depth from 0.5 ppt to 18.9 ppt, while the October 1987 values ranged from 0.1 ppt to 3.2 ppt. Salinity increased with depth during both surveys. The maximum range of values for one station at one time on October 1986 was 4.8 ppt to 18.9 ppt (Station 1) and for October 1987, 2.0 to 2.9 ppt. Over the duration of each hydrographic survey, minimum and maximum salinity values lagged behind the beginning of the flood and ebb cycles, respectively by one to two hours. Salt water continued to migrate upstream well into the ebb-tidal cycle (3 h and 15 min past flood slack) on October 1986 (see Station 1, Fig. 11).

The vertical distribution of salinity with depth reveals that the upper Kennebec River estuary is a partially mixed to moderately stratified estuary (Fig. 11). During slack water, salinity is vertically homogeneous with values less than 2 ppt. During the 1 October survey, salinity values were high enough along the bottom to result in a well-defined halocline. Salinity values for the October 1987 survey were not large enough to produce vertical stratification.

The water temperature for October 1986 and October 1987 surveys ranged from 15.2°C to 18°C on October 1986 and from 9.5°C to 11.2°C on October 1987. Temperature gradients were much less well-defined than salinity gradients and are not considered to contribute greatly to development of a pycnocline during the fall months.

5. Discussion

5.1. *Stratigraphy and sedimentation*

Knebel (1986), Belknap et al. (1986, 1987a,b, 1989), Kelley et al. (1992), and Barnhardt et al. (1995) used a depositional sequence approach, based

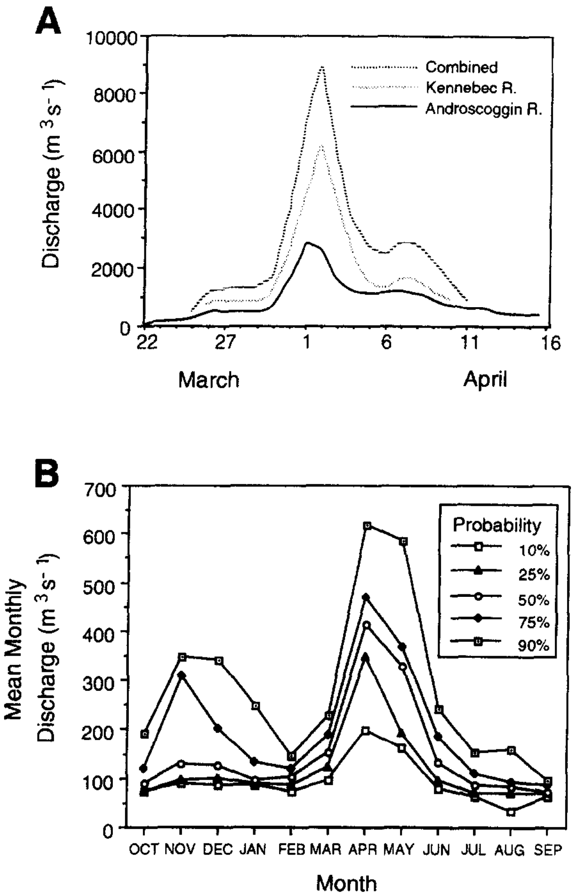

Fig. 10. (A) Stream hydrograph from the lower Kennebec River during record flood of April 1987. Flooding occurred when high temperatures melted record snowbank accumulations. After Stumpf and Goldschmidt (1992). (B) Seasonality analysis of discharge showing the monthly probability that the mean discharge will be less than the values indicated on the *y*-axis. Data from stream gauge I on the Androscoggin River (A_d = 8473 km^2). See Fig. 1B for location.

on the interpretation of high-resolution seismic data and core data, to provide evidence for five major events which controlled the geologic history of coastal Maine during the late Pleistocene and Holocene. From oldest to youngest, these include: (1) glaciation (>14 ka); (2) ice retreat and marine submergence (14 to 12.5 ka); (3) postglacial isostatic rebound and coastal emergence producing a lowstand shoreline (−55 m) and, in the vicinity of the Kennebec River, deposition of a paleodelta (11 to 10.5 ka); (4) rapid submergence (10.5 to 9 ka) followed by a slower rate of inundation under eustatic sea-level rise (9 to 5 ka); and (5) a modern slow transgression (5 ka to present).

As expected, subbottom data from the Kennebec River estuary show similar seismic facies as other estuaries along the west-central Maine coast. The basement consists of metasedimentary (schist) and plutonic (granite) Precambrian- to Paleozoic-aged rocks (Osberg et al., 1985). The irregular basement surface is overlain, in most places, by the blue clay of the Presumpscot Formation. The Presumpscot Formation is a glaciomarine clay deposited during an initial sea-level highstand (Bloom, 1960, 1963). The marine limit (and spatial extent of blue clay) in the Kennebec River estuary basin, as interpreted by topset–foreset contacts of glacial ice-contact deltas, is +130 m above present sea level (Thompson, 1982). In some reaches along the estuary, glacial diamict and stratified drift are interpreted to lie unconformably and discontinuously between the basement and the Presumpscot Formation. Our data and the bridge borings corroborate those of other studies conducted within nearby estuaries which show that diamict is not widely distributed within the estuaries (Belknap et al., 1989). Postglacial isostatic rebound resulted in sea-level (and base-level) lowering, coastal emergence (stage 3 above), and relatively planar erosion of the upper Presumpscot Formation surface.

During the initial stages of the Holocene transgression (stage 4 above), we hypothesize that the river mouth continued to receive coarse-grained sediment (sand and fine gravel), along with lesser amounts of silt and clay. During this phase, base level rose and the shoreline moved northward (landward). The locus of sedimentation, particularly the coarse fraction, migrated onshore as well, eventually depositing sands and fine gravels in the present estuary. Reworking and onshore transport of inner-shelf paleo-delta and paleo-shoreline deposits may have contributed additional sediment to the estuary (Belknap et al., 1989). A transgressive lag deposit (20 to 100 cm thick), consisting of sands and gravels, identified offshore of the present river may provide evidence of this process (Belknap et al., 1994; Barnhardt et al., 1995). Aggradation rates within the estuary decreased as the rate of sea-level rise slowed (stage 5 above) and the channel adjusted to the approximate present-day base level, riverine discharge, and tidal flow conditions.

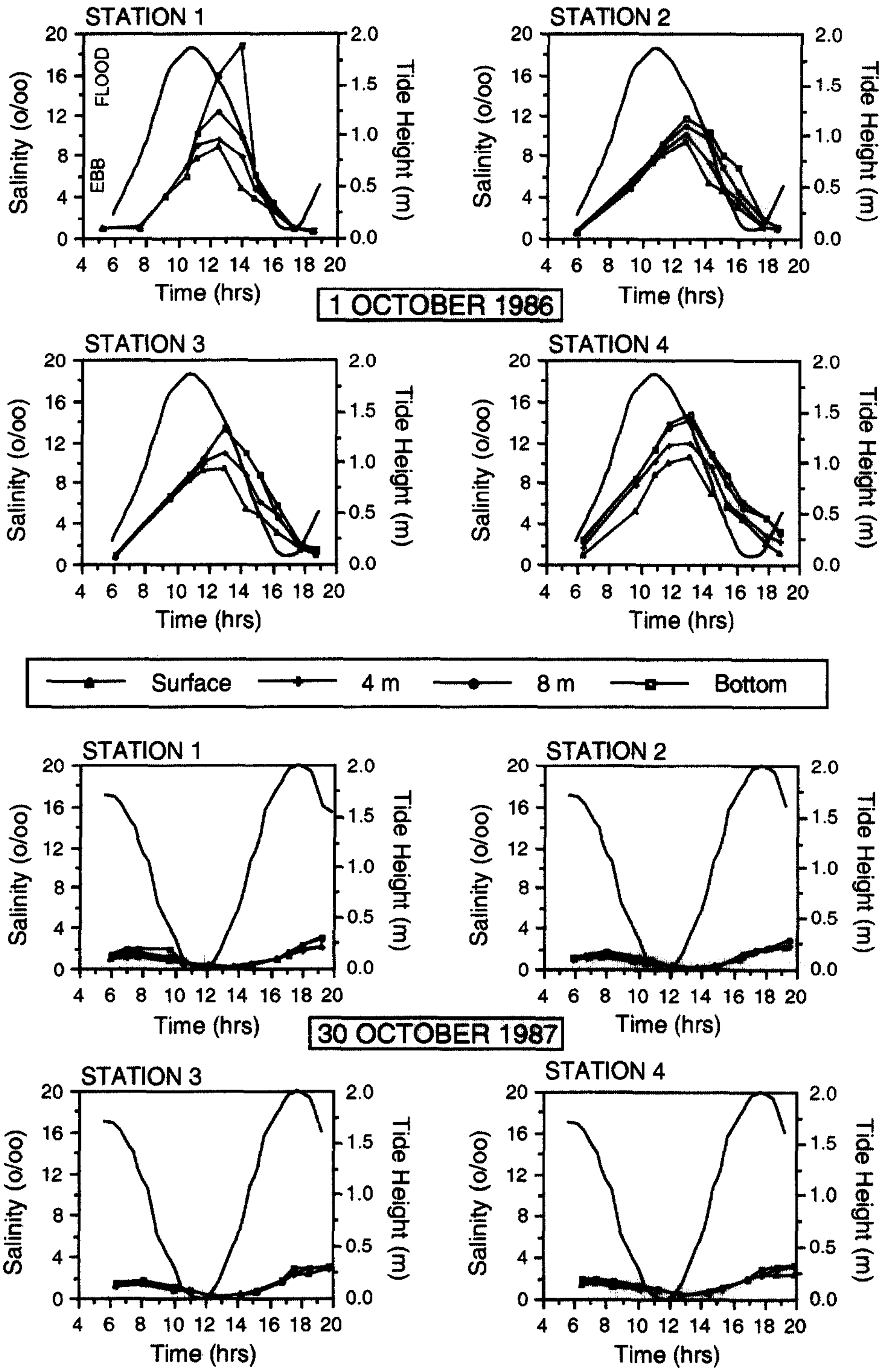

Fig. 11. Salinity values obtained during 1 October 1986 and 30 October 1987 hydrographic surveys as a function of tidal stage (see Fig. 3A for location of hydrographic survey stations). *Note:* (1) that during 1 October 1986 maximum salinity values occur nearly two hours after high slack for all stations and (2) the high saline content of the estuary waters. A similar lag existed during the 30 October 1987 survey, but with significantly lower maximum salinity values.

During the most recent period of low sea-level rise rates (stage 5 above), the estuarine unit has been subjected to reworking by reversing tidal currents and fresh-water discharge, as evidenced by the many discontinuous, discordant internal reflections within the upper Holocene-aged unit, and the hierarchical suite of superimposed bedforms molded on the unit's surface (the channel bottom). The internal seismic reflection patterns of this unit indicate that deposition occurred in a moderate- to high-energy environment and/or disruption of beds followed deposition (Sangree and Widmier, 1979). Larger-scale, more continuous unconformities have been preserved in the mid- to lower-portion of the Holocene-aged unit. Two of these features provide evidence for downstream sediment transport: (1) large-scale planar erosion surfaces, which emanate from the troughs of the transverse bars and extend upstream beneath one or more bars; and (2) smaller-scale, inclined bedding surfaces, which parallel and dip at the same angles as the modern-day bar lee slopes ($<6°$). We interpret the former features as bar 'migration ramps' (e.g., similar to trough migration basal unconformities; Allen, 1980) and the latter as bar slip-faces. Alternatively, the clinoform-like reflectors could be interpreted as accretion surfaces formed during an initial rapid fill of the estuary after Holocene flooding. However, the repetitive foreset bedding planes which parallel and have the same dip angle as the modern-day slip-faces, the bar migration ramps, and the morphology (asymmetry and relief) and orientation (ebb) of the modern bars provide evidence for downstream sediment transport following an initial period of embayment infilling. Moreover, long-term deposition as estuary fill would tend to smooth the bottom topography.

The Kennebec estuarine deposits are generally much coarser-grained than nearby estuarine sequences, which consist mostly of mud (Penobscot River, Knebel, 1986; Sheepscot River, Kelley et al., 1987; Damariscotta River, Belknap et al., 1994), due to a greater contribution of coarse-grained sediment supplied to this estuary throughout the Holocene. The shear stresses exerted on these sediments from tidal and fresh-water flows have produced a hierarchical suite of superimposed bedforms on the upper surface of the estuarine unit. The largest forms (transverse bars) are ebb-oriented and host a wide variety of bedform types (Fig. 4). The orientation of the transverse bars did not change over the two and a half year study period probably due to the greater amount of time that is necessary to move this large volume of sediment (Ashley, 1990). General sand wave patterns did not change over a period of two consecutive tidal cycles, but dominant orientations did reverse seasonally with ebb-orientations dominating in the late winter and spring months (Table 1). During all other sampled months, flood-oriented sand waves were dominant in the upper portion of the estuary, whereas the mouth of the estuary commonly exhibited ebb-oriented sand waves (FitzGerald and Fink, 1987). In some reaches, where bathymetric depressions, bedrock or sedimentary highs, and/or major changes in channel geometry occur, exceptions to these trends were observed.

5.2. Estuarine circulation

Temporal (monthly and annual) and spatial variations in density gradients influence the velocity profile and consequently, bedform migration (e.g., McCutcheon, 1981). The results from this study show that, similar to other Maine estuaries, the lower estuary (to 17 km north of the mouth) is stratified in the summer months when temperature and salinity gradients are maximum (Garside et al., 1978). However, during the spring months estuarine circulation is dominated by high fresh-water discharge conditions resulting in downstream flushing or diffusion of the salt wedge. The data from this study show that the salinity can increase five-fold with 71% less fresh-water discharge (150 $m^3\ s^{-1}$ on 1 October 1986 vs 211 $m^3\ s^{-1}$ on 30 October 1987) and a maximum temperature differential of only 1° on both dates (Fig. 11).

The complex channel geometry and bathymetry within the Kennebec River estuary may pose an impediment to the upstream or downstream migration of the salt wedge. For example, the lee slope of a transverse bar, a 90° bend in the channel, and an abrupt change in channel cross-sectional area are all located at Doubling Point, the downstream limit of our hydrography stations. The increased channel roughness and geometrical constraints to flow may cause boundary or diffusional effects (i.e., notice that ebb velocities > flood velocities for Station

4 bottom data only on 1 October 1986; Fig. 8). Since minimum salinity values for both survey dates approached zero, the upstream limit of salt encroachment occurs approximately 17 km from the mouth at Bath, Maine.

5.3. Net sediment transport trends and high discharge events

The Kennebec River estuary is similar to mid-latitude estuaries (e.g., Chesapeake Bay; Schubel and Pritchard, 1986) in that sedimentation, circulation, and stratification patterns are strongly influenced by variations in fresh-water discharge. Probability values (cumulative frequency distributions) of mean monthly discharge data for the Kennebec River–Androscoggin River watershed indicate that spring discharge values (April and May) exceed those of the summer and fall months by a factor of four to five and those of the early winter months by a factor of two to three (Fig. 10B). During years of large spring freshet events, these seasonal differences in fresh-water discharge are even more pronounced and can exceed summer discharge values by an order of magnitude.

Combining our remotely sensed data of bedform orientations with empirical flow data, dredge spoil (disposal site) bathymetric surveys, and existing theory we postulate that net sediment transport, via the en masse movement of subordinate bedforms (e.g., Allen, 1980), is towards the mouth of the estuary. Since the sediment transport rate is governed by the power function relationship (cubic or higher order) between sediment transport and friction (flow) velocity (e.g., Einstein, 1950; Bagnold, 1956; Engelund and Hansen, 1967; Sternberg, 1972; Yalin, 1977), net sediment transport is dominated by the stronger ebb-reinforced velocities associated with spring floods. Due to this relationship, a slightly stronger current velocity yields a dominant transport trend despite the current's short duration.

A plot of tidal range vs daily discharge for all survey dates shows: (1) that an apparent threshold exists in which sand waves maintain an ebb-orientation above discharge values of 225 m^3 s^{-1} to 325 m^3 s^{-1}; and (2) that sand wave orientation is independent of tidal range, except near the mouth where ebb-orientated sand waves abound during higher tidal range conditions (Fig. 12). The seasonality analysis reveals that the highest discharge values above the threshold occur during April and May and lower discharge values occur in November and December (Fig. 10B). Thus, downstream transport is ubiquitous during the late winter and spring months. This trend continues until mid-summer when fresh-water discharge diminishes and flood tidal currents predominate. A lesser degree of upstream (flood-directed) transport prevails during the fall and early winter until the transition to ebb-dominant transport occurs during the spring.

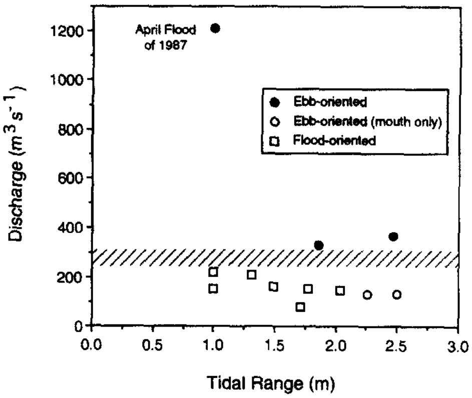

Fig. 12. Plot of tidal range and fresh-water discharge for all survey dates. □, correspond to surveys in which flood-oriented bedforms predominated in the Bath to Doubling Point region; •, correspond to flood-oriented bedforms in the Bath to Doubling Point region, but ebb-oriented bedforms near the mouth; o, correspond to ubiquitous ebb-oriented bedforms. This plot suggests that bedform orientation (net sediment transport) depends mostly on fresh-water discharge (ebb-reinforced flows) and the change from flood- to ebb-oriented bedforms occurs approximately 225 to 325 m^3 s^{-1} (hachured area) except near the mouth.

We hypothesize that the greatest amount of downstream movement of all large-scale bedforms (sand waves and transverse bars) in the Kennebec River estuary occurs during high-magnitude spring freshets and other large flood events (Fig. 13). For example, at peak flow during the April flood of 1987, the fresh-water discharge was twice that of the saltwater tidal prism over a tidal cycle. In fact, the fresh-water discharge supplanted the estuary's saltwater tidal prism from approximately 30 March through 3 April 1987. Additionally, peak discharges during the 1987 flood exceeded discharges during the fall and winter months by a factor of 50 to 60 (Fig. 10). During

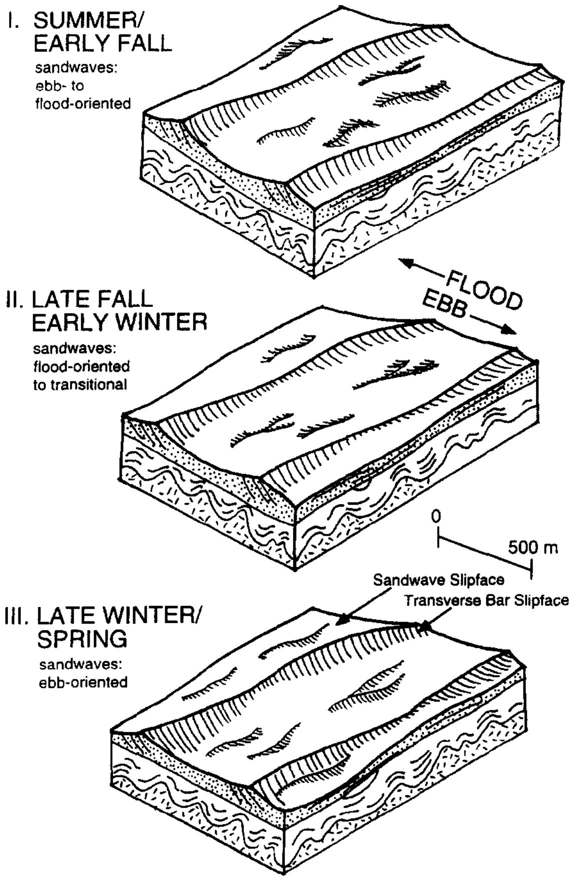

Fig. 13. Model depicting seasonal changes in bedform orientation and consequently, sediment transport directions. Note that while sand wave orientations change seasonally, transverse bar orientations remain ebb-oriented.

non-spring months, the estuarine sediment (stored in transverse bars) is reworked by reversing tidal flows. However, over the long-term, the net transport of coarse-grained material is downstream, supplying several beaches at the mouth of the Kennebec River with sediment.

The grain-size data from the river channel corroborate this finding. Coarse sand can be traced for more than 20 km from the Bath bridge seaward to the mouth of the river. Bottom samples recently collected along Popham Beach and seaward of the river mouth to a depth of 17 m indicate that medium to coarse sand circulates between the river, the adjacent beaches, and nearshore. Seaward of the river mouth, however, the grain size gradually decreases to fine sand (FitzGerald et al., in prep.). These trends point towards an upstream source for the coarse sand in the channel thalweg and not the offshore.

6. Conclusions

The subbottom geology of the Kennebec River estuary consists of three primary unconsolidated to semi-consolidated units, ranging in age from late Pleistocene to Holocene, overlying a Paleozoic-aged basement. From oldest to youngest, these units include glacial till and stratified drift, glacio-marine blue clay of the Presumpscot Formation, and Recent estuarine fill. The estuarine fill probably was deposited during the period of moderate transgression approximately 9 to 5 ka (Kelley et al., 1992; Barnhardt et al., 1995) when the system was equilibrating to a higher base level and to a larger degree of marine influence.

During modern slow transgression conditions, sediment transport within the estuary has shifted to an ebb-dominated system in which ebb-tidal currents, seasonally reinforced by fresh-water discharge events, dominate long-term bedload sediment transport direction and magnitude. Thus, the estuary is a sediment source for beaches along the west-central Maine coast.

The mode of sediment transport is via a hierarchical organization of bedforms in which large-scale, ebb-oriented transverse bars are reworked by subordinate bedforms (Fig. 13). During the spring, the subordinate forms are predominantly ebb-oriented, while flood-oriented forms predominate at other times of the year. Bedrock pinning points and channel geometry thresholds may control the positioning of large-scale transverse bars. Time scales responsible for the migration of the bars are longer than our three year study period.

During summer, fall and winter seasons, the estuary near Bath, Maine can be classified as a stratified to partially-mixed estuary. When fresh-water discharge is at a maximum during the spring, mixing is assumed to exist far downstream of Bath. Bath, Maine is the approximate upstream limit of salt incursion during moderate fresh-water discharge conditions.

Additional process data (concurrent hydrographic surveys at various locations throughout the estuary) are necessary to determine the spatial and temporal nature of energy inputs within the estuary. The combination of high-resolution bathymetric data and precision navigation techniques would be useful for determining bedform migration rates.

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