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Late Holocene Relative Sea-level Rise and the Geological Development of Tidal Marshes at Wells, Maine, U.S.A.

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ABSTRACT

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Tidal marshes along the Webhannet and Little Rivers of the Wells National Estuarine Research Reserve, Maine, U.S.A., are protected by barrier beaches and underlain by a relatively thick sequence of late Holocene estuarine and back-barrier sediment. More than 50 cores from the salt marshes at Wells were collected to evaluate geological and botanical development of the estuaries, and to determine the local, late Holocene rate of relative sea-level rise. New radiocarbon dates were employed to construct a local relative sea-level curve. Sea level rose at an estimated rate 0.8 mm/yr at 4,000 BP, slowed to 0.4 mm/yr at 2,000 BP, and 0.2 mm/yr at 1,000 BP. Modern tide gauges in Maine began to measure a contemporary rate of relative sea-level rise of around 1.0–2.0 mm/yr between 1940 and 1990, suggesting a much accelerated rate of rise analogous to mid-Holocene rates. During the mid-Holocene period of rapid sea-level rise, sediment collected in the deeply cut fluvial channels remaining from the lowstand of sea level. Interfluves of the estuaries, floored by glacigenic sediment, were covered by tidal flat deposits and the low marsh plant, *Spartina alterniflora.* Tidal channels migrated very slowly and remained confined to channels cut by rivers during the sea-level regression. By about 2,500 BP, the rate of sea-level rise slowed, and the marsh increasingly became dominated by the high marsh plant, *Spartina patens.* Rapid contemporary sea-level rise is associated with narrowing of the barrier beach and widening of tidal creeks, as the back-barrier and estuarine system adjusts to the new sea-level regime.

ADDITIONAL INDEX WORDS: *Botanical changes, Gulf of Maine, salt-marsh evolution, radiocarbon dating.*

INTRODUCTION

During the late Quaternary, northern New England (U.S.A.) experienced complex and extreme (> 200 m) excursions of relative sea level (BELKNAP *et al.,* 1987; KELLEY *et al.,* 1992b). In Maine (U.S.A.), a late-glacial marine incursion accompanied deglaciation as rising eustatic sea level overtook regional isostatic rebound of the land. This late-Pleistocene submergence was followed by a regression related to rapid isostatic rebound of the land, and finally by an extended Holocene transgression. The relative importance of eustatic sea-level rise, in comparison with local crustal motions as factors in the on-going transgression, has not been fully established and remains an important problem (ANDERSON *et al.,* 1984, 1989; BELKNAP *et al.,* 1987, 1989; GEHRELS and BELKNAP, 1993; REILINGER, 1987).

One way to evaluate the relative importance of land subsidence is to construct local and regional sea-level curves for the late Holocene. Studies involving conventional, radiocarbon-dated bulk salt-marsh peats (SHIPP, 1989; BELKNAP *et al.,* 1989) were unable to reject the hypothesis that rates of late Holocene sea-level rise were equivalent all along the coast of Maine, in part because variability in age-depth relations within individual marshes was large compared with the differences among marshes. A succeeding study, using accelerator mass spectrometry (AMS) ^{14}C dating, tide-gauge records and foraminiferal analyses, demonstrated a lack of crustal warping during at least the past 5,000 ^{14}C years (GEHRELS and BELKNAP, 1993). However, contemporary tide-gauge records (LYLES *et al.,* 1988) suggest that differential land movements are occurring. Therefore, one objective of this research is to evaluate the sources of variation in the local sea-level record at a single, intensively studied salt-marsh site (Wells, Maine, U.S.A.) and to reduce the uncertainties with addition of critical new data.

Some tidal marshes contain stratigraphic in-

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formation about the development of coastal environments as well as the rate of sea-level rise. REDFIELD (1965, 1972) built on LUCKE's (1934) earlier hypotheses on back-barrier evolution, and concluded that New England marsh systems show a prograding colonization of high marsh over low marsh over tidal flat, as well as high marsh over upland soils. Although this work has provided an important model for the origin of these systems, not all beaches in New England have formed by progradation of a simple spit. For example, recent work in Connecticut (GEHRELS and VAN DE PLASSCHE, 1991) and Massachusetts (FITZGERALD *et al.*, 1992; HILL and FITZGERALD, 1992) demonstrates that glacial deposits and paleo-valley systems shape contemporary coastal environments by providing sediment and loci for barrier beach development, respectively. BELKNAP and KRAFT (1985) and OERTEL *et al.* (1992) emphasize the influence of antecedent topography on back-barrier development in the mid-Atlantic region.

In northern New England, rock outcrops commonly interrupt the longshore movement of sand and tend to compartmentalize the beach systems, greatly complicating coastal development (KELLEY, 1987; DUFFY *et al.*, 1989). The great diversity of geologic settings along the coast of Maine has produced several different patterns of marsh development (KELLEY *et al.*, 1988). The different settings are also likely to have profound influences on the responses of these marshes to changing sea level. Therefore, a second objective of this research is to reconstruct the geological and botanical development of the Wells barrier beach and tidal-marsh system and to compare this reconstruction with existing models of beach and marsh development in New England.

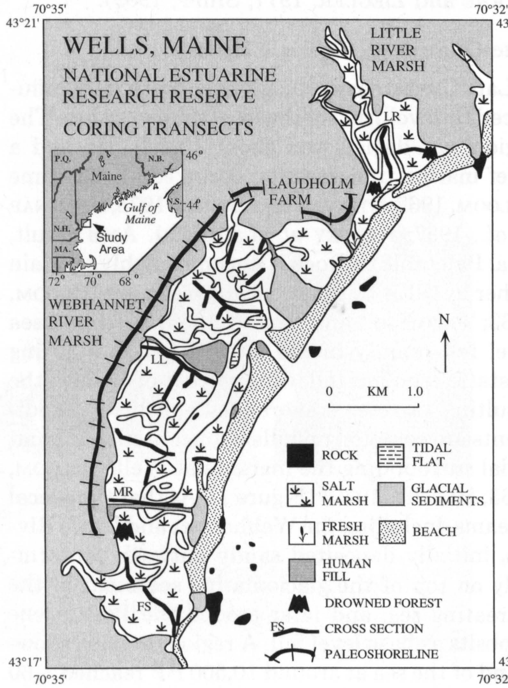

Figure 1. Location of core cross sections and geological environments and features in the Wells National Estuarine Research Reserve. Individual core loci are shown in cross sections in Figures 5 and 6.

STUDY AREA

Modern Physical Setting

The tidal marshes of Wells, Maine, including the Webhannet River marsh and the Little River marsh, lie within the Arcuate Embayments coastal compartment of the western Gulf of Maine (Figure 1) (KELLEY, 1987). This region, which extends from 50 km north to 100 km south of Wells, is characterized by many curved sandy barrier beaches (KELLEY, 1987; KELLEY *et al.*, 1988), which protect salt marshes from wave attack and are separated from one another by bedrock headlands. The marshes in this coastal compartment account for the largest proportion of tidal marsh area in Maine (JACOBSON *et al.*, 1987).

The Webhannet and Little River marshes are connected to the Gulf of Maine via tidal inlets that separate the barrier beaches. Well-mixed estuarine circulation results from tidal mixing (2.6 m mean tidal range) and low river discharge (each approximately 1 m^3/sec) (FITZGERALD *et al.*, 1989). Time/velocity asymmetries of the inlets lead to flood dominance and the development of large flood-tidal deltas (FITZGERALD *et al.*, 1989).

Today the Little and Webhannet Rivers apparently do not introduce appreciable volumes of sediment to their estuaries. During the late Holocene, however, shoreline erosion of bluffs of glacigenic sediment supplied sand and mud to the coastal zone. Although all such bluffs are presently protected by engineering structures, reworking of intertidal and subtidal exposures of glacigenic deposits may regularly provide some "new" sediment to the estuaries (HUSSEY, 1970; BYRNE and ZIEGLER, 1977; SHIPP, 1989).

Late-Quaternary Geologic History

Late Quaternary geologic events strongly influenced the evolution of the local tidal marshes. The region was deglaciated about 13,800 BP, and a brief marine transgression occurred at that time (BLOOM, 1963; STUIVER and BORNS, 1975; BELKNAP *et al.*, 1987; KELLEY *et al.*, 1992b). As a result, local Paleozoic bedrock is unconformably overlain either by till or by glaciomarine sediment (BLOOM, 1963; THOMPSON and BORNS, 1985). Regional sea level fell rapidly in the late Pleistocene during isostatic rebound following retreat of the ice; the resulting regressive shorelines and littoral sediments are common on hills of unconsolidated material surrounding the marshes at Wells (BLOOM, 1963; HUSSEY, 1970) (Figure 1). At that time local streams, including the Webhannet and Little Rivers, initially deposited sandy outwash conformably on top of the glaciomarine sediment in the retreating sea, and later eroded the Pleistocene deposits as base level fell. A regional relative lowstand of the sea at around 10,500 BP reached −60 m, from whence sea level subsequently rose to its present position (SHIPP *et al.*, 1991; KELLEY *et al.*, 1992b).

Surficial Features of the Tidal Marshes

The back-barrier region at Wells has been intertidal throughout its existence, and contained salt marshes for at least the past 4,000 years (BELKNAP *et al.*, 1989; SHIPP *et al.*, 1989). Today they are developmentally mature (*sensu* FREY and BASAN, 1985) and contain several distinct plant communities. Where not altered by humans, the forested upland margin is usually separated from the marsh by a 5 m to 20 m wide zone dominated by *Typha* sp., *Scirpus* sp., *Spartina pectinata* and, for approximately the past 10 years (based on personal observations), *Phragmites australis*. This brackish water wetland is characterized by many dead, standing trees along the upland edge, and by tree stumps in the marsh itself (Figure 2a and b). The trees there died presumably either from salt water reaching their roots, or simply from drowning by freshwater.

The higher high marsh that grades seaward from this brackish edge is characterized by a mosaic pattern of vegetation dominated by *Juncus gerardii* (JACOBSON and JACOBSON, 1989). Large quantities of plant detritus often accumulate here following very high tides, as do blocks of ice in the winter. Large salt pannes (local water-filled depressions) are common on the landward edge of the marsh (Figure 2b), as well as elsewhere. Erosive processes (*e.g.*, oxidation or ice-scour and plucking) along the margin of the pannes sometimes remove marsh peat and permit tree stumps to fall into the pannes. *Spartina alterniflora* and *Distichlis spicata* commonly colonize the panne margins, eventually invading the entire floors of the pannes once they drain (Figure 2b) (BERTNESS and ELLISON, 1987).

Spartina patens is the dominant plant of the extensive "high marsh meadow" (FREY and BASAN, 1985; NIXON, 1982; JACOBSON and JACOBSON, 1989). Other species in this zone include *Solidago sempervirens*, *Triglochin maritima*, *Scirpus americanus*, and *Puccinellia paupercula* (VADAS *et al.*, 1988). The marsh surface associated with *S. patens* is relatively flat, ranging from 1.2 m to 1.8 m National Geodetic Vertical Datum (NGVD) (mean = 1.51 m). Relief on the high marsh surface is associated with stream channels and blocks of ice-rafted peat, although much of the ice-rafted material there comprises inorganic debris rather than large peat blocks. The transition between the high and low marsh is usually abrupt and easily recognizable by the size difference between *S. patens* and the low marsh plant, *S. alterniflora* (Figure 3a). The mean elevation of the contact at the sites we surveyed is 1.35 m NGVD, close to mean high water, 1.43 m NGVD (HUBBARD, 1992) (Figure 4).

The low marsh is usually occupied solely by *S. alterniflora*, which ranges from 0.2 m to 1.4 m NGVD (Figure 4). Where our cores were collected, the mean elevation of thick stands of *S. alterniflora* is 1.17 m NGVD, relative to mean tidal level at 0.09 m NGVD (National Oceanic and Atmospheric Administration/National Ocean Survey, NOAA/NOS). The lower elevations represent areas where *S. alterniflora* is colonizing tidal flats (Figure 3b); whether it can survive there is unclear.

PREVIOUS PALEOENVIRONMENTAL RESEARCH AT WELLS

MCINTIRE and MORGAN (1964) were the first to collect cores from the Wells tidal marshes. They recognized glaciomarine sediment ("blue clay") and till at the bottom of many of their cores. On the basis of radiocarbon dates from basal peat

Figure 2. (a) Dead and dying *Pinus strobus* trees along the Little River higher high marsh; (b) Salt pannes with adjacent tree stumps (arrows) in the high marsh of the Webhannet River marsh.

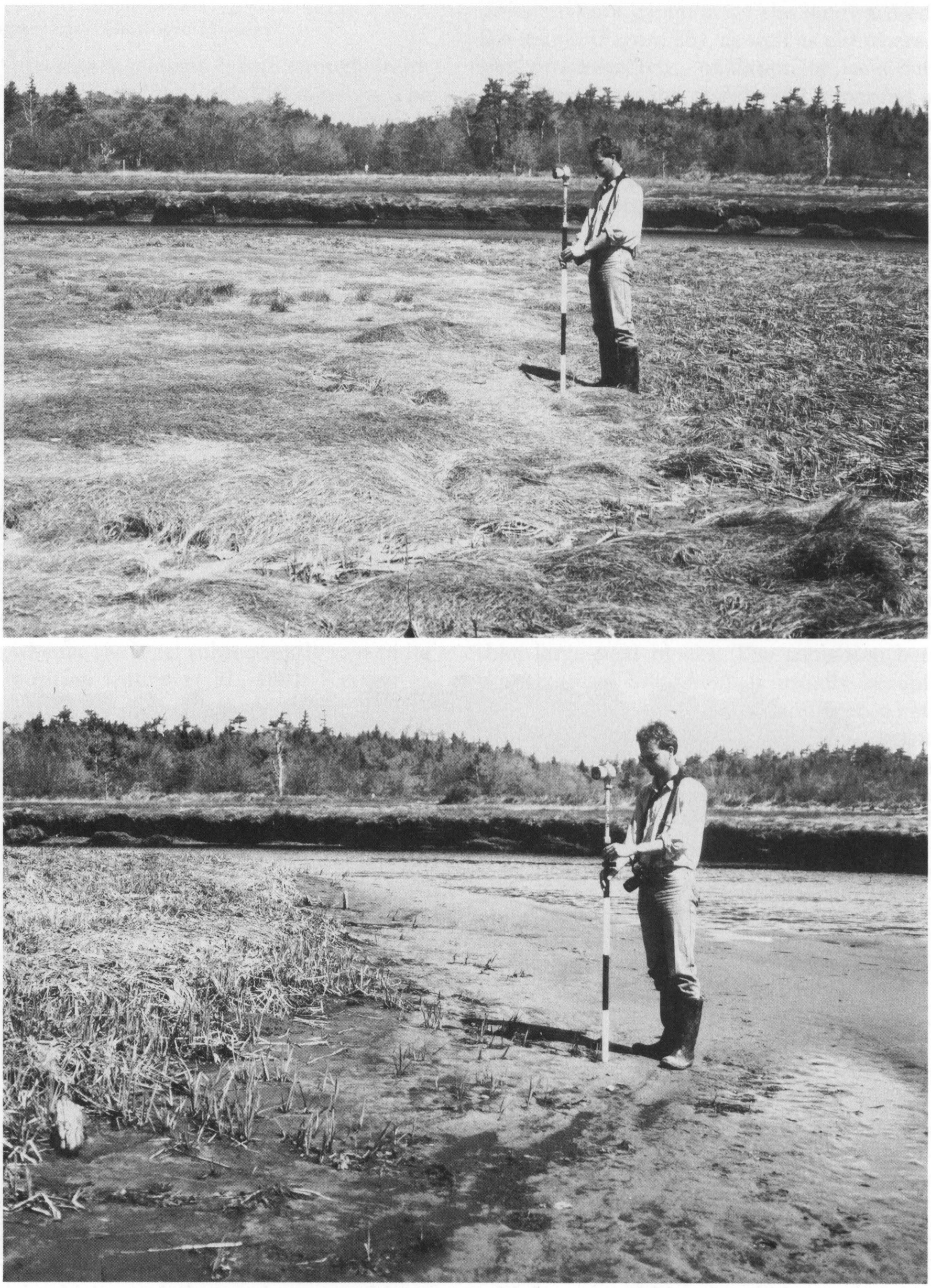

Figure 3. (a) Abrupt contact being surveyed between high (left) and low (right) marsh in the Little River marsh; (b) *Spartina alterniflora* propagating across a point bar of the Little River.

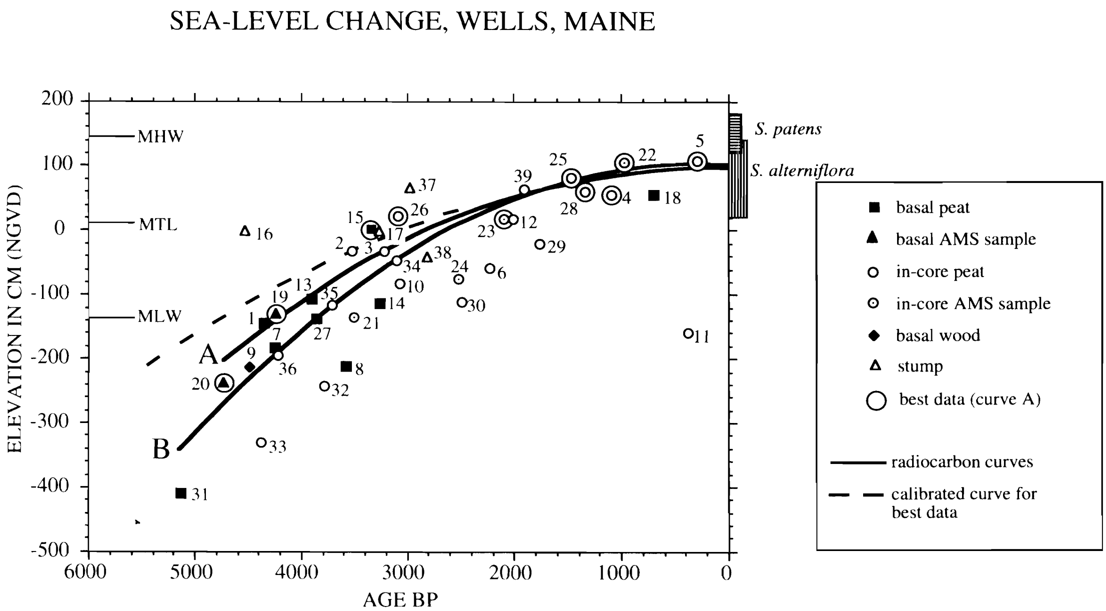

Figure 4. Sea-level data from the marshes of the Wells area. Curve A selectively incorporates only those samples deemed "best data", as described in the text. Curve B includes other basal high marsh radiocarbon dates. Dashed curve repositions Curve A following calibration of the radiocarbon dates (STUIVER and REIMER, 1986). Numbers correspond to samples in Table 1. The present, surveyed vertical range of the main salt marsh grasses is displayed in the upper right.

samples from Plum Island, Massachusetts, about 60 km south of Wells, they concluded that the rate of sea-level rise in the region slowed significantly between 4,000 BP and 2,500 BP. In Kennebunkport, Maine, less than 10 km north of the study area, a transitional fresh/salt-water peat at present sea level yielded a radiocarbon age of 1,900 BP, leading MCINTIRE and MORGAN (1964) to suggest that local sea level reached its present position then.

HUSSEY (1959) used radiocarbon dates on two drowned tree stumps from Wells Beach and on two from Kennebunk Beach (Table 1) to estimate that those beaches formed around 3,000 BP. Based on his observation of freshwater trees rooted in peat and covered by the present salt marsh, HUSSEY (1970) proposed that before 3,000 BP the site of the present Wells marsh was occupied by a marine lagoon, followed briefly by an emergent, forested upland, which was subsequently drowned by the sea shortly after 3,000 BP. TIMSON (1978) gathered cores and dated peat samples from the marsh, but no publication describes details of the work (Table 1).

BELKNAP *et al.* (1989) and SHIPP (1989) conducted a detailed stratigraphic study of the Webhannet Marsh as part of an investigation of regional sea-level changes in coastal Maine. On the basis of 12 radiocarbon dates from peat samples, they estimated that, from 5,100 BP to 1,500 BP, sea level at Wells rose 1.9 mm/yr; since 1,500 BP, sea level rose less than 0.5 mm/yr. The sea-level curve for Wells was statistically indistinguishable from those of other sites in Maine and northern Massachusetts (BELKNAP *et al.*, 1989; MCINTIRE and MORGAN, 1964; REDFIELD and RUBIN, 1962). When data from salt marshes throughout Maine were pooled, the best-fit sea-level curve rose at a rate of 1.9 mm/yr from 5,000 BP to 2,500 BP and at a rate of 0.71 mm/yr since 2,500 BP. All these dates were expressed as uncorrected ^{14}C years.

JACOBSON and JACOBSON (1989) examined the patterns of vegetation in the Little River marsh, among others, and demonstrated quantitatively that *S. alterniflora* dominates low marsh communities and is succeeded by *S. patens* and *Juncus gerardii* in the high marsh. JACOBSON (1988) compared maps of the Webhannet River Marsh

Table 1. *Information pertaining to the radiocarbon dates of samples from the Wells salt marshes. Numbers refer to plotted data in Figure 4 (samples 39-43 are not plotted due to uncertainties discussed in text). The elevation uses NGVD as a datum. LM = low marsh, HM = high marsh, HHM = higher high marsh, S.p. refers to a rhizome of* Spartina patens. *References are: a (this paper); b (Belknap* et al., *1989); c (Hussey, 1959), d (McIntire and Morgan, 1964), and e (Timson, 1978). Siev means the dated sample was sieved and coarser than 500 microns.*

Lab #	Graph #	Core	Elevation (meters)	Material	Age (^{14}C years)	Ref.
(PITT)						
0900	1	LR1	−1.42−−1.52	LM	4,335 ± 60	a
0898	2	LR1	−0.33−−0.39	LM	3,520 ± 60	a
0899	3	LR1 (Siev)	−1.72−−1.78	LM	3,215 ± 365	a
0897	4	LR1	0.55–0.48	HM	1,090 ± 50	a
0896	5	LR1	1.08–1.03	HM	300 ± 50	a
0906	6	FS1	−0.55−−0.6	LM	2,225 ± 60	a
0907	7	FS1	−1.79−−1.84	HHM	4,255 ± 55	a
0912	8	FS2	−2.08−−2.15	HHM	3,585 ± 60	a
0913	9	FS2	−2.12−−2.15	Wood	4,480 ± 60	a
0909	10	FS2	−0.80−−0.84	HM	3,065 ± 75	a
0910	11	FS2	−1.51−−1.55	HM	380 ± 55	a
0916	12	FS3	0.51–0.38	HM	2,010 ± 40	a
0917	13	FS3 (Siev)	−0.94−−1.1	HHM	3,900 ± 145	a
0918	14	FS3	−1.04−−1.19	HHM	3,265 ± 70	a
0920	15	FS4	0.05–0.02	HHM	3,340 ± 55	a
0962	16	Stump	*ca.* −1.4	Wood	4,535 ± 35	a
0963	17	Stump	*ca.* −1.4	Wood	3,260 ± 40	a
0902	18	LR2	0.54–0.4	LM	705 ± 165	a
(Arizona, AMS date)						
8208	19	FS10	−1.17	HM	4,235 ± 70	a
8209	20	FS5	−2.37	HM	4,735 ± 70	a
(Beta, AMS date)						
44064	21	FS2	−1.23	S.p.	3,510 ± 60	a
44061	22	FS2	1.11	S.p.	980 ± 55	a
44062	23	FS2	0.199	S.p.	2,100 ± 55	a
44063	24	FS2	−0.597	S.p.	2,520 ± 60	a
(SI)						
6618	25	MR1	0.87–0.79	HM	1,470 ± 55	b
6619	26	MR1	0.26–0.19	HM	3,080 ± 70	b
6620	27	MR7	−1.32−−1.4	HM	3,865 ± 55	b
6621	28	MR8	0.64–0.58	HM	1,345 ± 55	b
6622	29	MR8	−0.17−−0.23	LM	1,755 ± 55	b
6624	30	LL3	−1.06−−1.12	LM	2,495 ± 80	b
6623	31	LL2	−4.03−−4.10	HM	5,135 ± 70	b
6625	32	LL4	−2.37−−2.45	HM	3,780 ± 55	b
6626	33	LL5	−3.27−−3.34	HM	4,380 ± 55	b
6627	34	LL6	−0.42−−0.50	HM	3,105 ± 70	b
6628	35	LL6	−1.11−−1.17	HM	3,705 ± 50	b
6629	36	LL6	−1.98−−2.04	HHM	4,220 ± 60	b
(W)			(MHW)			
396	37	Stump	0.76	Wood	2,980 ± 180	c
508	38	Stump	1.83	Wood	2,810 ± 200	c
???	39	Marsh	−0.21	Peat	1,900 ± 105	d
???	40	Marsh	−0.23	Peat?	1,290 ???	e
???	41	Marsh	−0.58	Peat?	2,340 ???	e
???	42	Marsh	−0.83	Peat?	2,630 ???	e
???	43	Marsh	−1.07	Peat?	2,860 ???	e

dating from 1794 to 1984 A.D. and showed: (1) the salt marsh has become smaller due to erosion, and (2) with one exception the tidal creeks in the marsh have not moved appreciably during that time. The exception involves the stream just north of the large pile of dredge spoils (Figure 1). The pile is probably compressing the peat beneath it and causing the stream to migrate rapidly toward it.

METHODS

Marsh Stratigraphy

The general stratigraphy of the Webhannet and Little River marshes was established with fifty Eijkelkamp ("Dutch") hand auger cores collected along several transects across each marsh (Figure 1). Plant remains (*e.g.*, roots, rhizomes, and stem fragments) in these reconnaissance cores were identified by visual inspection in the field (BELKNAP *et al.*, 1989). Peat stratigraphy was later confirmed by detailed examination of pollen, macrofossils, water content, loss of weight on ignition, and foraminifera in the laboratory (KELLEY *et al.*, 1992a).

Sediment Cores

Using information gathered with the Dutch cores to select important sites within the marshes, six 7.6 cm diameter vibracores were collected in May 1990 by the method of LANESKY *et al.* (1979). Core sites were surveyed by infrared electronic theodolite and tied to the NGVD. The amount of sediment compaction that resulted from coring was determined by measuring the elevational difference between the marsh surface and the top of the peat cores prior to their extraction from the marsh (BELKNAP *et al.*, 1989). Compaction ranged from 10 cm to 56 cm and averaged 25 cm. Depth of samples within the vibracores was adjusted for compaction by establishing the depth of lithologic boundaries through comparison with non-compacted Dutch cores from the same site.

Radiocarbon Dating

Samples of either bulk sediment or individual plant macrofossils were dated by standard radiocarbon and AMS techniques, respectively (Table 1). Radiocarbon ages of bulk sediment were determined at the University of Pittsburgh Radiocarbon Laboratory. Samples were dried overnight at 100 °C and pretreated with 2 N NaOH and 2 N HCl prior to counting (BELKNAP *et al.*, 1987). Samples for AMS dating were removed from the center of the cores using tweezers, washed with distilled water, and dried overnight at 100 °C in aluminum foil. The samples were pretreated to eliminate carbonates and humic acids. Chemical pretreatments and target-material conversions were performed by Beta Analytic Inc., Miami, Florida; AMS measurements were made at Eidgenossische Technische Hochschule (ETH) University in Zurich. Detrital *S. alterniflora* fragments from the base of cores were similarly dated by AMS methods at the University of Arizona. Age estimates were based on the Libby half-life of 5,568 ± 30 years, with corrections based on secular tree-ring calibration.

CORE LOCATIONS AND DESCRIPTIONS

Little River Marsh

A series of Dutch cores was collected along the length of the central "peninsula" formed by the winding course of the Little River (LR), as well as across the marsh to the east (Figure 1) (KELLEY *et al.*, 1992a). At its northern end, the peninsula terminates in a point bar where the stream is eroding northward into marsh peat. The first four cores south from the tip of the peninsula each contain a sequence of sediments recording the infilling of the northward-migrating tidal creek (Figure 5a). The coring met with refusal in sandy-silty sediment identical to the present point bar, with overlying low marsh peat (*S. alterniflora*) followed by a mixed high/low marsh peat and capped by a high marsh peat (*S. patens*) (Figures 3a and b and 5a). A radiocarbon age of 705 ± 165 BP from the basal low marsh peat of LRVC-2 (#18, Table 1) establishes the beginning of modern peat accumulation at that site. With the core site approximately 30 m from where *S. alterniflora* is presently colonizing the point bar (Figure 3b), the mean rate of northward migration of the stream has ranged from 3.7–5.2 cm/yr.

Three cores near the southern end of the peninsula met refusal in glaciomarine sediment (Figure 5a). This distinctive blue-gray mud was unconformably overlain by freshwater peat and/or tidal creek silty sediment. Around 4,300 years ago (#1, Table 1), low marsh peat began to accumulate in the central peninsula, and continued for about 3,250 years. At 1,090 BP (#4, Table 1), high marsh peat succeeded the low marsh and persisted until the present. A sand layer, covering 300 year old peat (#5, Table 1) just beneath the marsh surface, extends through all the other cores in the southern part of the Little River marsh. It

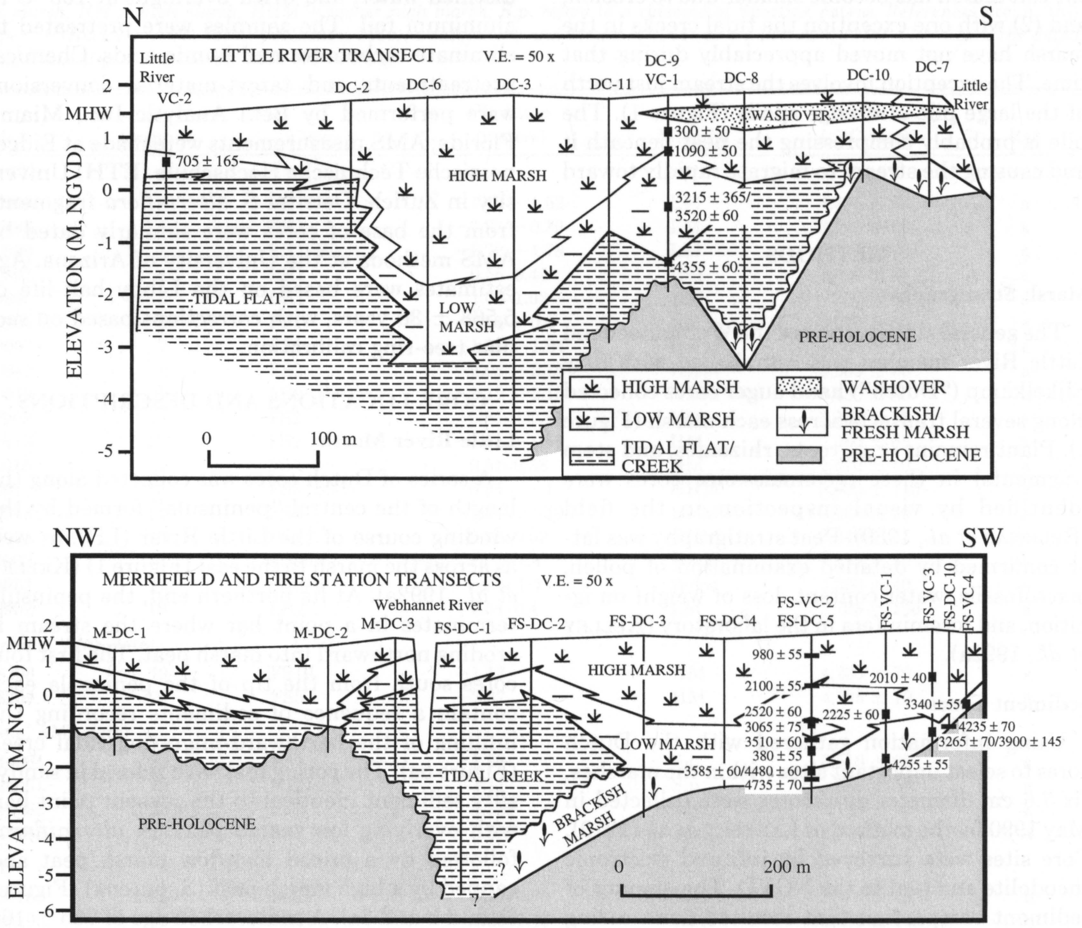

Figure 5. (a) Core transect from the Little River marsh (LR); (b) Core transects through the upper Webhannet River marsh (FS). The pre-Holocene is glaciomarine sediment along both transects. DC are dutch cores, VC are vibracores in Figures 5 and 6. Transects in Figures 5 and 6 are located in Figure 1.

thickens towards the inlet where it merges with storm overwash deposits and tidal delta sediments. Along the eroding bank near the tidal delta large tree stumps of *Pinus strobus,* white pine (identification by Richard Jagels, University of Maine College of Forest Resources), are exposed beneath the salt-marsh peat (Figure 1). Samples from two separate tree stumps, each about 2 meters below the high marsh surface, yielded radiocarbon ages of 4,535 ± 35 BP (#16) and 3,260 ± 40 BP (#17), respectively (Table 1).

To the east of the Little River, sediment cores revealed a channel filling similar to that near the tip of the peninsula (KELLEY *et al.,* 1992a). Refusal was generally met in silty sand, except near the eastern edge of the marsh where coarser, washover sand exists.

Webhannet River Marsh

Four traverses were made perpendicular to the Webhannet River, and connected by two traverses landward and sub-parallel to it (Figure 1). The southern traverse (Fire Station: FS) extends northwest-southeast across the upper estuary (Figure 1). Cores north of the Webhannet River met refusal in glaciomarine sediment (Figure 5b). This muddy sediment is overlain by tidal creek mud and/or low marsh peat, and capped by high marsh peat (Figure 5b). A layer of high marsh peat rests on the glaciomarine material along the eroding bank of the tidal stream.

On the south side of the Webhannet River a thick section of tidal creek muddy sand overlain by low, and then high, marsh peats records the infilling of the northward migrating stream. All the remaining cores from the traverse meet refusal in glaciomarine sediment at progressively shallower depths towards the southwest (Figure 5b). The glaciomarine contact is unconformably buried by peat with a "coffee-ground" texture (many unrecognizable plant fragments in a black, muddy matrix). The abundant wood and detrital plant fragments (occasionally, recognizable *Typha* sp. roots) were interpreted as wrackline accumulation in a brackish water, probably a higher high marsh setting, similar to today's estuarine margin (Figure 2). Toward the southeast, the base of this peat becomes younger from 4,735 BP (#20, Table 1) to 3,340 BP (#15, Table 1). Along the estuarine margin, high marsh peat rests on the higher high marsh material. Elsewhere the higher-high marsh unit is generally overlain by low marsh peat which grades upward into high marsh material. Several cores, however, contain alternating high and low marsh peat beneath surficial high marsh peat (Figure 5b). This probably indicates the nearby presence of a former tidal creek or salt pannes between 3,500 BP (#21, Table 1) and 2,500 BP (#24, Table 1).

Along the Mile Road traverse (Figures 1 and 6a) all cores west of the Webhannet River met refusal in glaciomarine sediment at increasing depths approaching the tidal channel. Thin muddy sandy tidal creek deposits rest unconformably on the glaciomarine material and are overlain by a thick deposit of freshwater peat with abundant wood fragments. The sequence is capped with high marsh peat except at the margin of the tidal creek and upland edge, where low marsh and freshwater peats, respectively, are accumulating. Most of the stratigraphic column on the east side of the tidal creek is composed of thick silty, sandy tidal-creek deposits, with relatively thin overlying deposits of peat.

Cores on the western end of the Lower Landing (LL) traverse (Figures 1 and 6b), as in the Mile Road Transect, penetrate only a freshwater (*Typha*) peat. The upper contact of that peat is progressively lower toward the east, and is buried by a high marsh peat that began to accumulate by 4,220 BP (#36, Table 1) (LLVC-6, Figure 6b). A sandy deposit, interpreted as latest Pleistocene regressive sediment, was at the bottom of all cores on the western part of the traverse (BELKNAP *et al.*, 1989), while sands that underlie the eastern portion of the line are interpreted as part of the tidal delta. On the eastern side of the traverse, tidal-delta sands are generally overlain by low marsh peat followed by high marsh peat (Figure 7b). In two locations, relatively old peats, 5,135 BP (#31) and 2,495 BP (#30, Table 1), possibly transported from elsewhere, were found within the tidal-delta sands.

Between the Mile Road and Lower Landing Road traverses, cores met refusal either in silty, sandy sediment near tidal creeks, or in glacigenic material in all interfluves. In the small marsh area behind the barrier beach all cores met refusal in glacial till, contrary to an earlier interpretation of the area as a tidal delta (FINK *et al.*, 1985). All cores from the central marsh area on both sides of the Webhannet River were covered by a high marsh peat over a low marsh peat.

Most cores north of the Lower Landing Road transect met refusal in glaciomarine sediment, which was buried by low marsh and then by high marsh peats. A diamicton (till) was encountered less than 2 m beneath the surface in two cores from the marsh just landward of the tidal inlet. Near the inlet itself, the flood-tidal delta is bordered by several meters of well-sorted sand, and then successively by low marsh and then high marsh peat.

SEA-LEVEL CHANGE

Sea-Level Indicators

Many efforts have been made to evaluate sea-level change in Wells, ME, and more radiocarbon dates are available from this area than most estuaries of comparable size (Figure 4; Table 1). The data reveal a general increase in elevation of the sea toward the present vertical range of *S. patens* and *S. alterniflora,* the principal salt-marsh plants. Not all data are plotted, however, and not all plotted points are valid sea-level indicators (Table 1). Five bulk peat samples (unplotted) yielded "modern" dates, or were less than 200 years old. Although the cores were not obviously disturbed, four came from within 10 m of a heavily used road and may have been contaminated by snow plowing or other human activity. However, one "modern" sample was from 1 m beneath the Little River marsh, which is undeveloped. Either vertical ac-

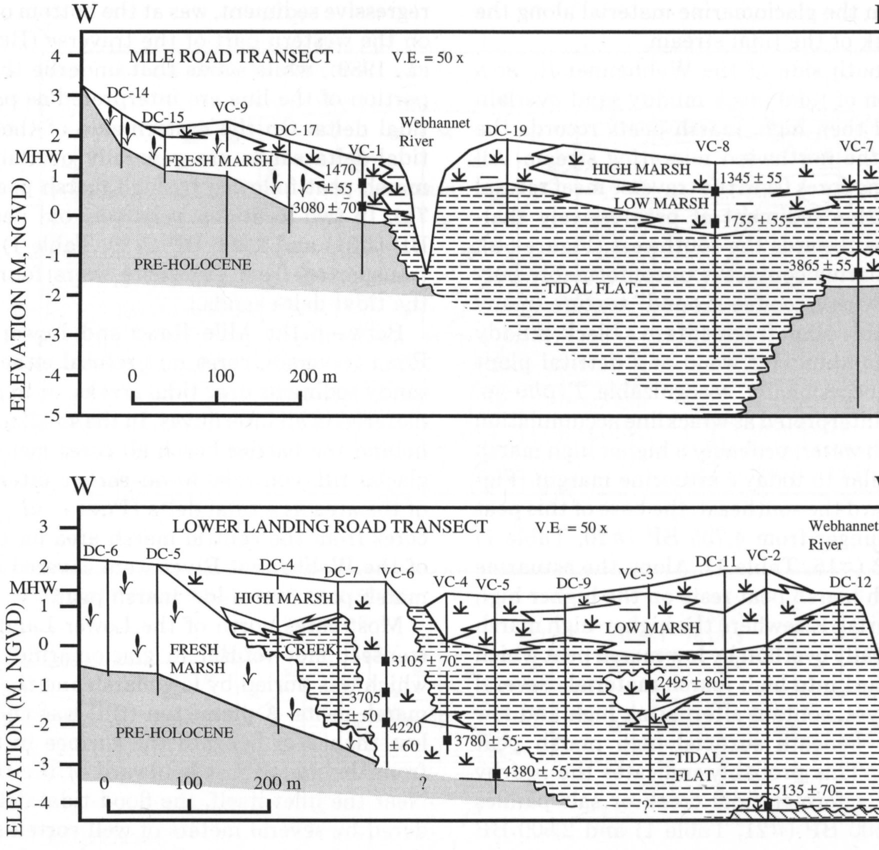

Figure 6. (a) Core transect from the Webhannet River marsh at Mile Road (MR). The pre-Holocene is glaciomarine sediment along this transect; (b) Core transect from the Webhannet River marsh at Lower Landing (LL). The pre-Holocene sediment appears to be regressive sand here.

cumulation of the marsh is rapid at this site, which is near a modern point bar, or young roots contaminated the sample. One sample from 2.7 meters beneath the surface of core FSVC-2 (#11) yielded a date of 380 BP, which is grossly inconsistent with ages of seven nearby samples within the core.

Despite their obvious advantages for radiocarbon dating, wood samples are probably the least reliable indicators of past sea-level position in this type of setting (Figure 4; Table 1). Trees may perish for a variety of reasons that are unrelated to rising sea level (Figure 2a and b) (Kaye and Barghoorn, 1964), and their deep root systems confound attempts to determine an elevation that can be tied reliably to a precise tidal datum. For example, the two trees from the same approximate depth near the Little River (#16 and #17, Table 1) yielded ages over a thousand years apart, without overlapping standard deviations (Figure 4). Thus, we do not consider these or other dates from stumps (#37 and #38; Figure 4; Table 1) as valid sea-level points. Furthermore, fragments of wood are easily transported and redeposited by tidal action. Although a fresh piece of wood could be buried at the high tide line, one cannot know this at the time of sample collection. A salt-marsh peat sample from the Fire Station traverse appears almost a thousand years younger than a wood fragment from the same depth in the same core (#8 and #9, Table 1; Figure 4), although this wood fragment date corresponds better with an AMS date from nearby (#20, Figure 4; Table 1).

Low marsh peats are also relatively poor indicators of past sea level. The surveyed vertical range of *S. alterniflora* in the Wells area is 1.3 m (Figure 4), similar to values obtained elsewhere in the region (HUGHES, 1975; FEFER and SHETTIG, 1980; MCKEE and PATRICK, 1988). The plant has a deep root system which further extends its vertical range. Today, the vertical distribution of *S. alterniflora* is significantly correlated with tidal range along the East Coast of the United States (MCKEE and PATRICK, 1988). Because the tidal range in the Gulf of Maine has changed over the past several thousand years (SCOTT and GREENBERG, 1983), the relationship of *S. alterniflora* to mean sea level may have changed as well. Owing to their ambiguous relationship to sea level, we do not use bulk low marsh peat samples (#1, 2, 3, 6, 18, 29, 30; Figure 4; Table 1) to construct our sea-level curve. We have used the AMS method to date individual stems of *S. alterniflora* lying flat on the paleo-wrackline near the basal unconformity (#19, 20; Figure 4; Table 1). Dead mats of *S. alterniflora* commonly accumulate at the higher high water mark, which was further verified by foraminiferal as well as stratigraphic considerations (GEHRELS, 1994). Because detrital salt-marsh plants do not persist very long when exposed, such samples are ideal sea-level indicators.

The best material for determining past sea-level positions was previously assumed to be basal high or higher high marsh peat resting on an uncompactable substrate (KAYE and BARGHOORN, 1964). Seven of the dated samples from Wells are basal high marsh or higher high marsh peats (#7, 8, 13, 14, 15, 27, 31; Table 1, Figure 4). Because there are more than 1,000 years of scatter between those samples older than 3,000 BP due to humic acid contamination (BELKNAP *et al.,* 1989; GEHRELS and BELKNAP, 1992), we prefer the two AMS dates (#19, 20; discussed above) from the same interval to construct a sea-level curve (Curve B includes these basal dates for comparison; Figure 4). We do use one basal high marsh sample (#15) where there is no better material at the same elevation to serve as a proxy.

All of the best-suited samples are greater than 3,000 years old and do not, by themselves, yield a useful sea-level curve. For the sea-level curve to be extended beyond the range of the basal peat samples, dates on non-basal high marsh peat younger than 3,000 BP (#4, 5, 22, 25, 26, 28; Figure 4; Table 1) must also be incorporated into the regression curve (Figure 4). We consider this to be acceptable, as did BELKNAP *et al.* (1989), because compaction within the Webhannet and Little River Marsh peats is minimal compared with that in the highly organic peats discussed by KAYE and BARGHOORN (1964). This is partly because Maine "peats" often contain less than 10% organic sediment (KELLEY *et al.,* 1992a; WOOD, 1990), contrary to general textbook descriptions of New England salt-marsh peats (FREY and BASAN, 1985, p. 244). In addition, the inorganic matter incorporated into northern New England high marshes (often transported by ice rafting) commonly contains sand and gravel, which further reduces its compactability (WOOD *et al.,* 1989; WOOD, 1990). Finally, we speculate that during the 2–3 month period of ice cover each year, the weight of the ice (which routinely exceeds 20 cm in thickness) squeezes the marsh, accelerating early compaction and reducing later changes in volume.

Sea-Level Reconstruction

The sea-level curve derived from the radiocarbon ages of the best high marsh peat samples rises steeply in the middle Holocene and flattens out toward the present (curve A, Figure 4). At 4,000 BP sea-level rose at a rate of 1.2 mm/yr, but slowed to 0.5 mm/yr at 2,000 BP and 0.2 mm/yr at 1,000 BP. The new curve differs from that of BELKNAP *et al.* (1989) in its assessment of the mid-Holocene (1.91 mm/yr between 5,000 and 2,500 BP, BELKNAP *et al.,* 1989), and late Holocene rates of sea-level rise (0.71 mm/yr between 2,500 BP and present, BELKNAP *et al.,* 1989) as well as in its overall shape. Those differences arose partly because we surveyed the core sites to a consistent datum (NGVD) and selectively introduced the new data points for this study. We also chose a quadratic fit to the data rather than 2 linear segments. When the ^{14}C data are calibrated (STUIVER and REIMER, 1986), the curve flattens out even more in the middle Holocene with rates of rise from 0.8 mm/yr at 4,000 BP (Figure 4).

The new sea-level curve for Wells also differs from that constructed for Connecticut by VAN DE PLASSCHE (1991), who inferred significant variations in the rate of late Holocene sea-level rise that we could not identify in Wells. We have assumed that such fluctuations are masked by random errors, which are best minimized by a least-squares model. Errors may also arise from contamination of dated material. For example, wood fragments that accumulate at the high tide wrackline may be significantly older than adjacent, living, high marsh plants. Peat samples containing different concentrations of wood fragments from the same stratigraphic level would thus yield different ages. In contrast, fine material may be younger than the true stratigraphic age. Root penetration introduces varying amounts of younger material that cannot always be seen or removed before radiocarbon dating on bulk samples. This is also true of humic matter, which is introduced through groundwater, and cannot be completely removed by the radiocarbon laboratory (R. STUCKENRATH, University of Pittsburgh, personal communication). GEHRELS and BELKNAP (1992) found that individual grass fragments dated by AMS were as much as 30% older than equivalent bulk peat dates.

Despite these differences in approach, we concur with VAN DE PLASSCHE (1991) and others (OLDALE, 1985; REDFIELD and RUBIN, 1962; KAYE and BARGHOORN, 1964) that the rate of sea-level rise slowed during the late Holocene. The environmental stability created by the relatively long, recent stillstand permitted colonization of most of the Webhannet and Little River marshes by high marsh plants. These mature marshes (*sensu* FREY and BASAN, 1985) expanded in the estuaries until the recent (historical) increase in the rate of sea-level rise. In the past century each of the three mareographs within 150 km of Wells (Boston, Seavey Island, and Portland: LYLES *et al.*, 1988) recorded rates of sea-level rise greater than even the rapid middle-Holocene rates reported here (0.8 mm/yr at 4,000 BP on the calibrated curve; Figure 4). This has apparently resulted in the observed narrowing of the barrier beaches, widening of the tidal creeks, and accompanying erosion of significant acreage of salt marsh (JACOBSON, 1988). The numerous salt pannes on the high marsh surface may also be a response to the prolonged period of sea-level rise.

PALEOGEOGRAPHIC RECONSTRUCTIONS

Development of the Webhannet and Little River Estuaries

The Webhannet and Little River marshes have changed greatly during the past 4,000 years (Figure 7a). At approximately 4,000 BP, tidal flats or channels occupied much of the Webhannet Estuary with high marsh, higher high marsh or freshwater marshes apparently colonizing only the landward margins where glaciomarine sediment provided a gentle transition to the upland (Figure 7a). It is possible that more high marsh existed

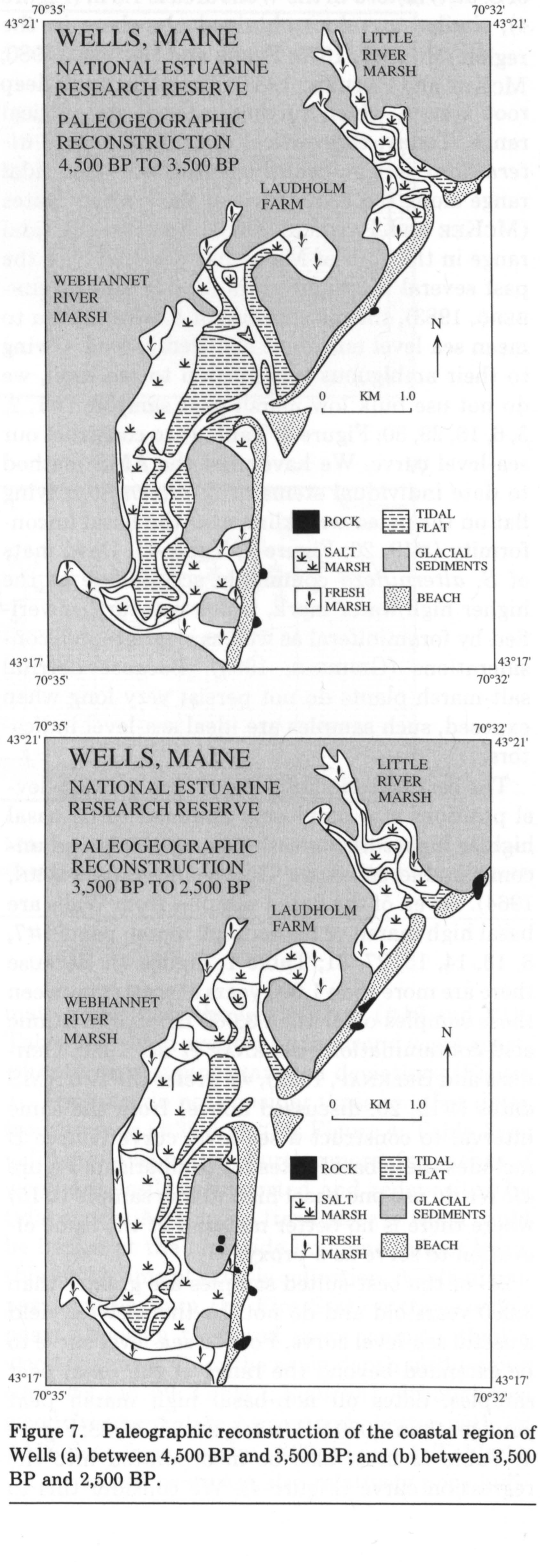

Figure 7. Paleographic reconstruction of the coastal region of Wells (a) between 4,500 BP and 3,500 BP; and (b) between 3,500 BP and 2,500 BP.

at this time, but that peat was removed by subsequent meandering of tidal creeks. The 5135 BP date from a possibly allochthonous block of high marsh peat (LLVC-2, #31, Table 1) indicates that *S. patens* peats did exist somewhere in the Webhannet estuary during the middle Holocene. Relatively extensive freshwater marshes are inferred at the head of many of the tributaries to the Webhannet, although no dates were obtained from sites where that could be confirmed. Most salt-marsh peat accumulating adjacent to the tidal flats and channels was low marsh, and this forms the lowest salt marsh peat in most cores. Low marsh was also accumulating in the central Little River marsh at this time. The beaches were in a more seaward location at 4,000 BP, and more extensive deposits of glacial till are shown covering today's rocky islands and shoals, as suggested by Hussey (1970).

By about 3,000 BP, beaches had retreated landward and the eroding glacigenic deposits had provided sufficient material to allow colonizing low marsh to outpace the rate of rising sea level (0.6 mm/yr), and thus to cover most of the tidal flats (Figure 7b). High marsh succeeded the low marsh in the central Webhannet Estuary and apparently continued to transgress across gentle upland slopes at the central and southern margins of the embayment. Isolated hills of till were slowly drowned, both in the northern Webhannet estuary and along the back side of the Wells barrier. The area of freshwater marshes was reduced as the region influenced by salt water approached the knickpoints of smaller streams related to rock outcrops and regressive shorelines. We made no observations to support the presence of stumps in the Webhannet Estuary at this time, as suggested by Hussey (1970).

The two streams occupying the Little River Estuary probably began to move into their present meander patterns by around 2,000 BP (Figure 8a). Their rate of meander was retarded by the relatively erosion-resistant glaciomarine sediment in their path. Sea-level rise had slowed to 0.4 mm/yr, and high marsh peat began to spread over the low marsh as well as the upland.

By 1,000 BP sea level had nearly reached its present location, and its rate of rise was 0.2 mm/yr (Figure 8b). The beaches had probably accreted seaward of the remaining glacigenic bluff deposits, thereby reducing the availability of inorganic sediment. High marsh peat continued to cap the low marsh deposits, and freshwater plants were

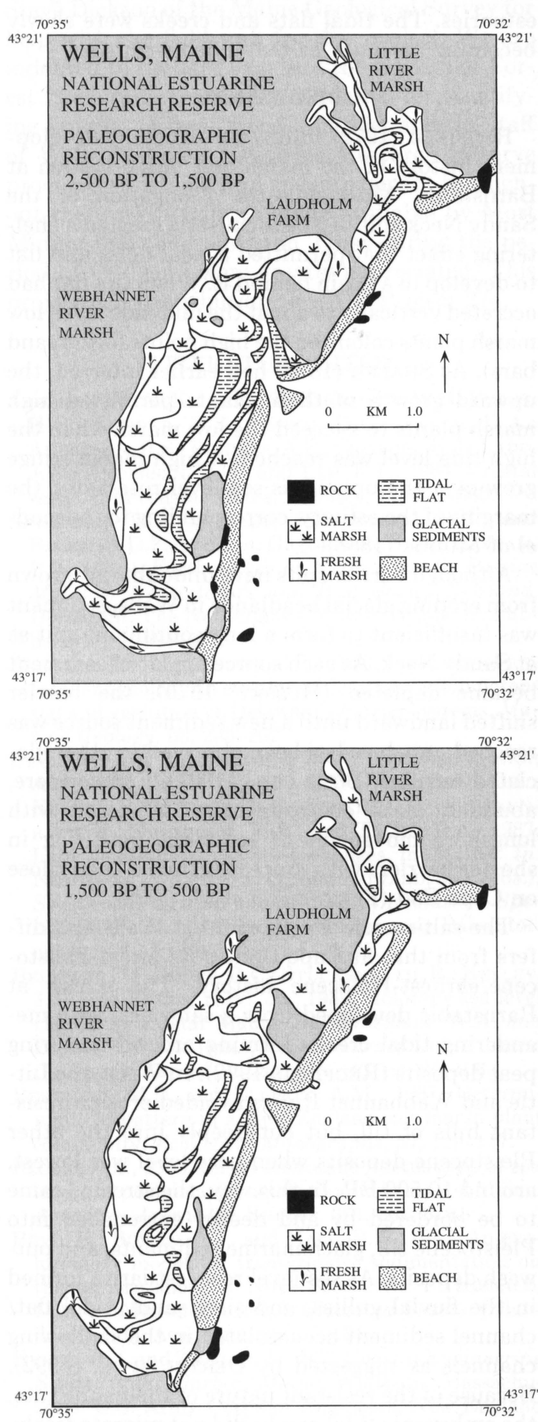

Figure 8. Paleographic reconstruction of the coastal region of Wells (a) between 2,500 BP and 1,500 BP; and (b) between 1,500 BP and 500 BP.

restricted to occasional narrow patches along the estuaries. The tidal flats and creeks were slowly becoming narrower as they filled with peat.

DISCUSSION

REDFIELD (1972) inferred a different development history of the marsh and beach system at Barnstable, Massachusetts. Elongation of the Sandy Neck spit in Massachusetts created a sheltering effect that permitted a tidal delta and flat to develop in a basin behind it. When the flat had accreted vertically to about the mid-tide level, low marsh plants colonized the high points (often sand bars). As SHALER (1885) had earlier inferred, the upward growth of these plants permitted high marsh plants to succeed the low marsh when the high tide level was reached. A high marsh fringe grew over the upland as sea level rose along the margins of the estuary, corresponding to the model of MUDGE (1858).

Although barrier spits have undoubtedly grown from eroding glacial headlands in Wells, sediment was insufficient to form a long continuous spit as at Sandy Neck. As each source of glacial sediment became depleted (HUSSEY, 1970), the barrier shifted landward until a new sediment source was reached, much as has been observed in other glaciated terrains (BOYD *et al.*, 1987). Furthermore, abundant rock outcrops have interfered with longshore movement of sediment, resulting in shorter barriers with more tidal inlets than those on Cape Cod (FITZGERALD *et al.*, 1989).

The salt-marsh development at Wells also differs from that at Cape Cod in its latest-Pleistocene/earliest-Holocene history. The marsh at Barnstable developed from a sand flat, with meandering tidal creeks forming around maturing peat deposits (REDFIELD, 1972). In Wells, the Little and Webhannet Rivers avoided erosion-resistant hills of till, but cut deeply into the other Pleistocene deposits when base level was lowest, around 10,500 BP. In this way, the streams came to be bordered by and deeply entrenched into Pleistocene till, glaciomarine sediments, and outwash deposits. As sea level rose, estuaries formed in the fluvial gullies, and silty sandy tidal flat/channel sediment accumulated in the shallowing channels as suggested by OERTEL *et al.* (1992). Because of the resistant nature of their substrate, the embryonic tidal creeks did not migrate freely, and tidal creek deposits are still entirely confined to the narrow meander belts of the regressive rivers. This is consistent with JACOBSON'S (1988) finding that the streams have moved little in Wells during the past 300 years. Relatively broad divides supported by glacigenic deposits still exist between the channel deposits; the thickest deposits of peat have accumulated over low divides of glaciomarine sediment. Where more elevated hills of till occur, such as between the Little and Webhannet River estuaries, along the Wells barrier beach, or within the northern portion of the Webhannet River marsh, they have retarded the rate of marsh expansion and confined salt marshes to areas underlain by gentler topography.

On a smaller scale, dominant vegetation is controlled by the pace of sea-level rise. In more than half of the cores, tidal flat/tidal channel deposits are the oldest Holocene unit recognized, and most of these were from the central area of the estuaries (Figure 6). Only along the fringe of the estuaries were freshwater, low and high marsh peats found at the base of the Holocene deposits. Thus, when sea level was rising rapidly (Figures 4 and 7), tidal flats and low marshes were probably the most abundant intertidal environments. *S. alterniflora,* which dominates the low marsh, is better suited to thrive during periods of rapid sea-level rise than are high marsh plants, because it tolerates longer periods of tidal inundation (MCKEE and PATRICK, 1988). This in turn leads to greater concentrations of compaction-resistant, substrate-building inorganic sediment (KELLEY *et al.*, 1992a). *S. alterniflora* also survives across a greater range of depths than *S. patens* (Figure 4), and so can endure in years when the sediment-accumulation rate falls below the rate of sea-level rise.

By about 1,000 BP, when the rate of sea-level rise had apparently slowed, high marsh succeeded low marsh across most of the estuaries. The high marsh plant community has dominated the estuaries from then until the present, apparently because sea level has risen sufficiently slowly to allow even more accumulation of organic-rich sediment. This accumulation keeps most of the marsh surface within the narrow vertical depth range tolerated by *S. patens* (Figure 4).

Within the marsh, we found no widespread evidence of oscillations in the rate of sea-level rise. The vertical sequences of marsh types within the estuaries are typical of what has often been described from the region previously (*e.g.*, REDFIELD, 1972; BELKNAP *et al.*, 1989). Low marsh colonized tidal flats in the central portion of the embayment and was later covered by high marsh. Tidal channel or high marsh followed freshwater deposits or soils along the margins of the embayments. In some instances, low marsh appeared directly above freshwater or high marsh sediment (FSVC-2, FSDC-4; Figure 5b), but such cases were neither widespread nor at a consistent depth or age. For those, migration of a nearby tidal creek followed by colonization by *S. alterniflora* provides a simpler explanation than accelerated sea-level rise. On a scale of centimeters, beds of *S. alterniflora* were occasionally found in a high marsh setting, or beds of *S. patens* occurred in a low marsh setting. Ice-rafting, salt-panne formation, or tidal creek meandering likely explain these relatively rare findings.

CONCLUSIONS

(1) AMS-dating of individual stems of *Spartina* deposited in a higher high water setting and surveying core elevations to a common benchmark improves the quality of sea-level change curves by reducing the variance associated with conventional dating of bulk salt marsh peat samples and assumed common elevations to all salt marsh surfaces.

(2) The rate of relative sea-level rise slowed until recent years and permitted marshes in northern New England to become developmentally mature. The recent, more rapid rise in sea level, at least in this century, appears responsible for rejuvenation of the marshes. This rejuvenation of the marsh system may be expressed as a change from high to low salt marsh, a change from low marsh to salt pannes, or through a widening of tidal creeks and transformation from salt marsh to tidal flat.

(3) Antecedent topography plays an important role in back-barrier marsh development in glaciated regions. Erosion-resistant glacial deposits confined rivers to relatively narrow channels during times of lowered sea level. As sea level has risen, interfluves of glacigenic sediment have continued to act as barriers to stream migration, and controlled the thickness of Holocene sediment.

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