Evolution of paraglacial coasts in response to changes in fluvial sediment supply

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Abstract: Paraglacial coastal systems are formed on or proximal to formerly ice-covered terrain from sediments derived directly or indirectly from glaciation. This manuscript reviews the roles of tectonic controls, glacial advances and retreats, sea-level changes and coastal processes on sediment production, delivery and re-distribution along the paraglacial Gulf of Maine coast (USA & Canada). Beaches and barriers along this coast are characterized by lithological heterogeneity and spatially variable sediment textures. They are found primarily at the mouths of estuaries, particularly those associated with the Kennebec/Androscoggin, Saco and Merrimack rivers. The formation of these barrier systems is directly attributable to the availability of sediments produced through the glacial erosion of plutons within the river basins and locally along the coast. Multiple post-glacial phases of sea-level rise and fall drove the redistribution of these sediments across the modern coastal lowland and shallow inner shelf. Most important for the formation of barrier systems was the *paraglacial sand maximum*, a period of relative sea-level fall and enhanced fluvial sand delivery 3–5 thousand years following glacial retreat. These sediments were reworked landward during the subsequent marine transgression and combined with younger sediments derived from the river basins to form the modern barrier and backbarrier systems. Today, reduced fluvial sediment loads due to natural depletion of glacially liberated sediment as well as anthropogenic modifications to the barriers and river systems combine with an increasing rate of relative sea-level rise to intensify beach erosion. These changes may also accelerate disintegration of backbarrier marshes and eventually force the Gulf of Maine barriers to return to states of rapid landward migration.

Glaciations perturb large parts of the global landscape to a greater degree and over a shorter time period than any other surface process. Gross sedimentation rates associated with glaciations are much higher than during interglacial periods (Broecker *et al.* 1958) and the mean rate of sediment delivery from ice sheets is an order of magnitude higher than from fluvial activity in some of the largest river systems in the world (Dowdeswell *et al.* 2010). Glaciers leave behind large volumes of easily erodible unconsolidated sediment that are subsequently redistributed by non-glacial earth-surface processes (Ryder 1971; Church & Ryder 1972).

Modern coasts located within the sphere of influence of these formerly glaciated terrains are known as *paraglacial coasts* (Forbes & Syvitski 1994). The concept of a paraglacial environment was first developed in terrestrial settings to describe the non-glacial processes that were directly conditioned by glaciation and occurred literally 'beyond the glacier', in environments located around and within the margins of glaciation (Ryder 1971; Church & Ryder 1972). It has since come to include not only the processes, but also the land systems, landscapes and sediment accumulations that are directly conditioned by glaciation and deglaciation. Paraglacial environments are unstable or metastable systems experiencing transient responses to a variety of non-glacial processes, acting over a number of spatial (metres to hundreds of kilometres) and temporal (years to thousands of years) scales to drive the systems toward recovery from glaciation (Ballantyne 2002a; Hewitt *et al.* 2002; Knight & Harrison 2009; Slaymaker 2009). Paraglacial environments are found in all regions of the globe that underwent or were directly influenced by glaciation during the Pleistocene (Fig. 1) (Mercier 2009).

Paraglacial coasts include erosional and depositional coastal landforms such as fjords and coarse-clastic barriers as well as formerly glaciated shelves (Forbes & Syvitski 1994). Such coasts fringe more than 30% of Northern Hemisphere continental shelves: they are common throughout northern North America, northern Eurasia, and Greenland (Forbes & Syvitski 1994; Forbes *et al.* 1995; Forbes 2005). In the Southern Hemisphere, the southern tip of South America and ice-free parts of Antarctica are prominent examples (Fig. 1). These coasts retain the recognizable influence of glacigenic sediments or morphologies (Forbes & Syvitski 1994). They are distinguished from many of their non-paraglacial counterparts by a unique combination of: (1) glacially overprinted landforms (such as fjords and drumlin fields); (2) nourishment by heterogeneous sand and gravel sources; (3) variable rates of sediment supply governed by substrate erodibility and impacted by terrestrial and marine processes; and (4) a high degree of compartmentalization (Forbes & Syvitski 1994).

Barrier islands and beaches are common along most paraglacial coasts. They are generally small (10^2 – 10^4 m long; 10^4–10^9 m^3 sediment volume) and relatively isolated (FitzGerald & van Heteren 1999). Many of the most prominent and continuous paraglacial barriers flank the mouths of rivers draining glaciated terrains. The goal of this paper is to provide an idealized stratigraphic and process framework that describes the unique features of barrier formation over periods ranging from hundreds to thousands of years in such river-associated paraglacial settings. This is accomplished by comparing three river-associated barriers along the paraglacial coast of the Gulf of Maine (GoM), located in the northeast United States (USA) and southeast Canada (Fig. 2), and by contrasting these systems with river-associated barriers formed in other paraglacial and non-glaciated settings. We review the dominant controls common to the formation of these coastal systems (underlying geology; glacial advance and retreat, relative sea-level changes, sediment redistribution by fluvial, coastal and marine processes) and their affiliated deposits. Detailed examples of these processes and deposits are drawn from the evolutionary histories of the barriers associated with the Kennebec/Androscoggin rivers (Kennebec barrier chain), the Saco River (Saco Bay barrier system) and the Merrimack River (Merrimack Embayment barrier chain) (Table 1; Fig. 3a). These details and comparisons are then used to refine and expand upon earlier concepts of the *paraglacial period* by reviewing the role of various contributions of glacial, paraglacial and nonglacial sediment sources to the development of the GoM barriers. Finally, we assess the future of the GoM barrier systems given climate change and human interference in the natural sediment-supply pathways.

River-associated paraglacial coasts

Most barriers located along paraglacial coasts are mainland-attached spits or bayhead barriers, formed through a cyclic pattern of formation and destruction as finite local sediment sources (such as drumlins, till bluffs and outwash) are periodically made available as relative sea-level (RSL) rises and shorelines migrate landward across formerly glaciated terrain (Boyd *et al.* 1987; Forbes & Taylor 1987; Duffy *et al.* 1989). They are generally composed of coarse-grained sand and gravel; coarse-clastic beaches up to boulder size are common as fines are progressively removed from the system (Orford & Carter 1985; Carter *et al.* 1989; Forbes *et al.* 1995). Even tidal flats, with sediment derived from erosion of glacial deposits, commonly contain a substantial gravel component.

Longer and more voluminous paraglacial barriers directly nourished by glacial sediment are generally confined to zones proximal to the maximum ice limit of the Last Glacial Maximum. On the coasts of outer Cape Cod, Massachusetts, and southern Long Island, New York, for example, RSL rise and coastal processes have reworked ample, easily eroded sediment from terminal moraines and expansive outwash (sandur) plains since deglaciation. Here, most barriers are moderately long (2–12 km), 200 m to > 1 km wide, 5–25 m thick and composed primarily of sand (Rampino & Sanders 1981; FitzGerald *et al.* 1994; FitzGerald & van Heteren 1999).

By contrast, paraglacial coasts located at the mouth of major rivers have received sediment not only from the erosion and direct reworking of local glacial deposits, but also from sediments delivered by rivers from the erosion of both glacigenic and paraglacial, upstream sources (Forbes & Syvitski 1994; Ballantyne 2002a; Forbes 2005). In New England, such river-associated paraglacial coasts have been influenced by late-Pleistocene and Holocene RSL changes. During many millennia, the sea has eroded and smoothed landforms as they passed through the nearshore zone, reworking previously deposited glacial, paraglacial and fluvial sediments into swash-aligned barrier island which front extensive backbarrier marshes and tidal flats (Kelley 1987; Hein *et al.* 2012). These barriers are composed chiefly of sand with a minor gravel component.

Coastal evolution in the Gulf of Maine: Dominant controls and common deposits

The GoM covers an area of ca. 93,000 km^2 from Cape Cod to the southern tip of Nova Scotia (Fig. 2). It is fringed by almost every type of paraglacial coast and has been ice free for 12–17 thousand years depending on position along a north-south gradient. Thus, its shore presents an ideal location to investigate glacial and paraglacial coastal evolution, and to study coastal environments and deposits formed at various stages of post-glacial recovery. The major controls on coastal evolution in the GoM are antecedent (bedrock-governed) structure, glaciation-dominated sediment generation, river-dominated sediment supply, RSL change, and sediment redistribution by marine and coastal processes.

Structural Control

The bedrock in the western GoM and on the adjacent land is dominated by granite, granitic gneiss and metasedimentary and metavolcanic rocks ranging from Precambrian to Middle Palaeozoic in age (Osberg *et al.* 1985; Lyons *et al.* 1997; Robinson & Kapo 2003). The major GoM rivers flow to the south-southeast, generally across the structural grain of bedrock, except locally where erosion of weaker strata by both fluvial and glacial scouring controls brief strike-aligned courses (Kelley 1987).

Structural controls related to Palaeozoic tectonism have played a dominant role in shaping the variability between barrier systems in the western GoM. For example, the southward-facing Kennebec/Androscoggin coastline is largely protected from northeast waves by bedrock headlands, resulting in the development of swash-aligned barriers (Buynevich 2001). By contrast, the Merrimack River drains into structural lowlands, providing ample accommodation for the development of the associated northeast-facing (drift-aligned) barriers, but less protection from dominant waves (Goldsmith 1991; FitzGerald *et al.* 2005). Faulting and folding patterns control the configuration of individual coastal compartments, and the gradients of the coastal lowland and proximal continental shelf control accommodation available for backbarrier and offshore deposits. Barriers throughout the GoM are commonly pinned to bedrock promontories, with abundant granitic plutons in southwest Maine serving as headlands anchoring and protecting beaches and marshes (Kelley 1987). Tidal inlets are typically situated in drowned bedrock-controlled river valleys (FitzGerald *et al.* 2002).

Advancing and receding ice sheets

Quaternary glaciations resulted in significant sediment and bedrock erosion across the GoM and in extensive scouring of bedrock-controlled fluvial channels. Physical and chemical weathering of intrusive granitic (with minor gabbro and granodiorities) plutons common to inland regions throughout the GoM (Fig. 3b), as well as glacial excavation of saprolite, generated much of the sand-rich sediment that was later reworked into glacial and paraglacial deposits (Hanson &

Caldwell 1989; Thompson *et al.* 1989; FitzGerald *et al.* 2005). Ice sheets of the most recent glaciations, the Illinoian and Wisconsinan, left behind non-stratified glacigenic sediments throughout the GoM and in adjacent New England. These take two dominant forms: (1) drumlins, which have a patchy, clustered distribution and were deposited in part during the Illinoian glaciation; and (2) coarse-grained till of late Wisconsinan age (Stone *et al.* 2006). This latter deposit drapes bedrock surfaces throughout the region as a thin veneer. Contemporaneously, meltwater streams drove the accumulation of extensive, quartz-feldspar-rich, stratified ice-contact sediments throughout the region, both under and in front of the ice sheet. The distribution of sandy eskers, outwash plains and fans, and coarse sandy ice-marginal deltas (Fig. 3b) reflect the ice-sheet extent and recession (FitzGerald *et al.* 2005).

The Laurentide Ice Sheet of the Wisconsinan glaciation reached its maximum extent, beyond the southern boundary of the GoM (Fig. 1a), between 28.0 and 23.7 thousand calendar years before present (ka) (Balco *et al.* 2002). As it then receded rapidly northward between *ca.* 17 and 15 ka (Borns *et al.* 2004), rising RSL in the GoM submerged isostatically depressed areas immediately upon deglaciation (Fig. 4) (Bloom 1963). This submergence culminated in a highstand of RSL several tens of metres above modern mean sea level (MSL) (Fig. 4), prelude to a subsequent set of complex RSL changes that served to redistribute the sediments produced by the glaciers.

Relative sea-level change

Late-Pleistocene and Holocene RSL changes (Fig. 4) resulted from the combined forcings of global eustatic sea-level rise and regional glacio- and hydro-isostatic adjustments. These RSL changes were largely similar across the GoM coast, varying only in amplitude and timing (Fig. 4).

While near its maximum extent, the mass of the Laurentide Ice Sheet depressed the lithosphere, including that of the GoM region. The outward displacement of the underlying asthenosphere formed a peripheral crustal forebulge 10s to 100s of kilometres beyond the glacial front (Daly 1934; Barnhardt *et al.* 1995). The ice sheet receded northward as climate warmed, exiting present-day Massachusetts by *ca* 17–16 ka and coastal Maine by *ca* 15 ka (Borns *et al.* 2004). Global eustatic sea level rose rapidly during this period. The continuous isostatic depression of the crust below contemporary MSL due to delayed rebound resulted in immediate marine flooding of land in the GoM (Bloom 1963). Isostatic rebound halted this process and the maximum marine limit was reached at 31–33 m above modern MSL in Massachusetts (Fig. 4) (Stone & Peper 1982; Oldale *et al.* 1983; Ridge 2004; Stone *et al.* 2004) and at 70–75 m above MSL in coastal Maine (Fig. 4) (Thompson *et al.* 1989; Kelley *et al.* 1992).

Continued isostatic rebound resulted in rapid RSL fall as regional uplift outpaced global eustatic sea-level rise. RSL stabilised as rebound decelerated and temporarily matched the rate of eustatic sea-level rise. This produced a relative marine lowstand across the GoM that ranged from *ca.* -41 m MSL at 13–14 ka (Oldale *et al.* 1993) in northern Massachusetts to -60 m MSL at 12.5 ka in central coastal Maine (Barnhardt *et al.* 1997) and -65 m MSL at 11.5 ka along eastern Nova Scotia (Fig. 4) (Stea *et al.* 1994).

Following this regional relative lowstand, RSL rose very rapidly (*ca.* 40 mm/yr) in coastal Maine for about 1000 years as eustatic sea-level rise greatly outpaced isostatic rebound. RSL rise then slowed abruptly to only 1–1.5 mm/yr at 11.5 ka as the forebulge that migrated north through coastal Maine collapsed (Barnhardt *et al.* 1995). This so-called slowstand (Fig. 4)

(Kelley *et al.* 2010, 2013) along the Maine coast lasted until 7.5 ka. No such detailed information is available for the southern GoM, where sparse post-lowstand data indicate that sea level rose at a time-averaged rate of *ca.* 4 mm/yr between 13.5 and 6 ka.

RSL rise in the southern GoM gradually slowed to less than 2 mm/yr by 4–5 ka (Fig. 4). By contrast, coastal Maine saw one final period of relatively rapid (*ca.* 7.5 mm/yr) RSL rise between 7.5 and 5.5 ka, followed by 5 ka of much slower change (less than 1 mm/yr) to its modern vertical position (Fig. 4) (Barnhardt *et al.* 1995; Kelley *et al.* 2010, 2013).

Deposits associated with changing sea levels

In governing the changing extent of the GoM, RSL rise and fall set the stage for large-scale erosion and redistribution of glacigenic sediment across the modern coastal zone and adjacent shallow shelf, generating a range of coastal, marine and glaciomarine landforms and deposits (Table 2; Fig. 5).

Following the Last Glacial Maximum, receding ice sheets across the GoM left behind extensive stratified and non-stratified deposits, including sandy eskers, outwash plains and fans, coarse sandy ice-marginal deltas and till (Stone *et al.* 2006). In the GoM region, this latter deposit is an unsorted mixture of mud, sand and gravel in various proportions and lithological compositions, depending on the material eroded and on the influence of eroding and transporting processes.

Flooding of land immediately following deglaciation resulted in the deposition of glaciomarine silt and clay across the modern shelf and adjacent terrestrial environments, occasionally extending several hundred kilometres inland from the modern shoreline (Thompson

& Borns 1985). Although this deposit smoothed antecedent topography, it provided limited sediment for the later development of barriers and beaches. Glaciomarine silt and clay was generally deposited in environments of low wave energy along the coast and in river estuaries. Its strongly micaceous composition indicates derivation from glacially eroded metamorphic rocks (Kelley 1989).

Falling RSL following the 17–15 ka highstand forced rapid shoreline progradation. Sediments delivered to the contemporary coast were reworked cross- and along-shore by coastal processes, forming sandy parasequences and regressive deltas that reflect the seaward migration of the coastal zone (Oldale *et al.* 1983; Barnhardt *et al.* 1997; Belknap *et al.* 2002; Kelley *et al.* 2003). Coastal and fluvial processes modified the landscape and redistributed sediments across discrete segments of emergent glaciomarine plains. Discontinuous, disparate remnants of regressive beaches and spits and fluvial terraces have been identified between the highstand and modern coastlines along much of GoM (Retelle & Weddle 2001). Near several river mouths, these regressive deposits formed extensive and thick strandplains and braidplain deltas (braided deltaic plain) as ample fluvially derived sand and fine gravel was efficiently reworked alongshore by waves and tides (Fig. 3b; Table 2). The Sanford-Kennebunk braidplain delta for example, located upstream of the Mousam River near Wells, Maine (Fig. 3b), has a surface area of 125 km^2, a thickness of 5–14 m. Its volume of 1.5 x 10^9 m^3 (Tary *et al.* 2001) exceeds that of all individual modern-day GoM barriers. The Brunswick braidplain delta, located at the former confluence of the Kennebec and Androscoggin rivers, covers an area of 25 km^2 and is 5–15 m thick. It gradually steps-down in elevation from west to east, reflecting underlying seaward-dipping bedrock gradually exposed by falling RSL (Crider 1998). Similar regressive units have also been found on the shallow shelves proximal to several river mouths. For example, off the mouth of the Merrimack River, abundant coarse sediment produced a 10-km-wide strandplain that parallels the shore for 16 km and is 4–15 m thick (Barnhardt *et al.* 2009). Its upper surface is marked by a lag deposit of coarse sand and fine gravel, formed during the later Holocene transgression (Hein *et al.* 2013).

Deceleration and subsequent cessation of RSL fall led to the deposition of lowstand delta lobes at the mouths of several major rivers discharging into the contemporary GoM, most notably the Merrimack and Kennebec/Androscoggin rivers (Oldale *et al.* 1983; Barnhardt *et al.* 1997). No equivalent deltaic deposits have been uncovered offshore the Piscataqua (New Hampshire / Maine border), St. John (New Brunswick) and Saco rivers. Some deltaic sediments were deposited in Penobscot Bay (Fig. 3a) before *ca.* 9 ka. However, sediment supply was largely cut off as the Penobscot River lost its competence to transport sand when isostatic uplift in its headwaters led to a drainage-divide shift and associated loss of water and sediment to the Kennebec River (Kelley *et al.* 2011). As much as 10 m of Holocene mud now covers this pseudo-palaeodelta (Belknap *et al.* 2002; Kelley *et al.* 2011). Eroded eastward-facing fossil shoreline features abound in Saco Bay, but these are very thin (less than 1 m) (Kelley *et al.* 2003).

The period of rapid RSL rise following the lowstand led to the surficial erosion of regressive and lowstand deposits and to the formation of a thin (< 1 m) transgressive sand-and-gravel unit on newly formed shelves. Muddy sands started to accumulate during the slowstand period in Maine (Fig. 4) to form wedge-shaped estuarine units over eroded glaciomarine mud, thickening toward the modern shoreline and showing large regional differences (Barnhardt *et al.* 1997). Off the Kennebec River, the estuarine unit is up to 10 m thick. In Saco Bay, a similar but muddier and much thinner (up to several metres) unit marked by an intertidal to shallow-subtidal fauna extends from the present shoreline out ~2 km offshore (Kelley *et al.* 2005; D. Barber, personal communication).

The broad time-transgressive sand sheets that occupy the shallow shelves between the lowstand and modern shorelines evolved from sediment not only reworked from early transgressive barriers and intertidal / supratidal sand shoals (Oldale 1985; Oldale *et al.* 1993; FitzGerald *et al.* 1994) but also contributed by direct fluvial input and by wave-driven erosion of glacial and regressive deposits exposed at the seabed. As the rate of RSL rise decreased, coastal processes associated with slowly retrograding shorelines drove sediments from coastal and river sources ever farther onshore, eventually forming proto-barriers (McIntire & Morgan 1964; FitzGerald *et al.* 1994; van Heteren 1996; Buynevich 2001; Buynevich & FitzGerald 2003). In many areas, sand and silt derived from fluvial and nearshore sources were deposited in backbarrier lagoons, tidal inlets and channels, and flood-tidal deltas as these proto-barriers lengthened and widened to their modern dimensions (van Heteren 1996; Buynevich 2001; Hein *et al.* 2012).

Freshwater and brackish marsh deposits initially formed at the leading edge of the transgression (McIntire & Morgan 1964). Most observations suggest a rapid transition from barren tidal flats to well-developed high marsh around or after 4 ka (Oldale 1989; FitzGerald *et al.* 1994; Kelley *et al.* 1995b). The cause and timing of salt-marsh expansion in New England estuaries has generally been attributed to a late-Holocene decrease in the rate of RSL rise. However, local settings and processes must not be neglected: salt-marsh development behind the Saco barriers, for example, is hypothesized to have coincided with a reduction in the width of the palaeolagoon, when an inferred palaeobarrier system was replaced by a more landward series of barriers. Previous to this event, a combination of rapid RSL rise and strong winds across wide lagoons had maintained open-water conditions (Van Heteren 1996). Today, these salt-marsh systems are dominated by *Spartina alterniflora*, *Spartina patens*, black rush, *Juncus gerardii* and the shrub Iva frutescens.

Riverine sediment redistribution

Rivers draining *ca.* 160,000 km^2 of land discharge into the GoM (Fig. 2; Table 3). Their drainage basins contain voluminous glaciofluvial and glaciomarine deposits. Together, the rivers annually deliver *ca.* 950 x 10^6 m^3 of freshwater and more than a million cubic metres of suspended sediment (Kelley *et al.* 1995a). Directly or indirectly, via temporary storage on the inner continental shelf, sediment input from these rivers has supplied nearly all of the sediment for the development of most barrier and backbarrier systems of the GoM (FitzGerald *et al.* 2005; Kelley *et al.* 2005; Hein *et al.* 2012); without them barriers in the GoM would be rare or absent, even in areas marked by extensive inland sand sources (FitzGerald *et al.* 2002). Sediment feeding the Saco and Merrimack barriers was dominantly provided by the rivers themselves (Kelley *et al.* 2003; FitzGerald *et al.* 2005; Hein *et al.* 2012). Elsewhere, marine erosion and reworking of shelf and post-glacial fluvial deposits, and, to a lesser extent, coastal erosion of drumlins (Chute & Nichols 1941; Dougherty *et al.* 2004; Hein *et al.* 2012), have also played a role.

Modern fluvial sediment delivery is episodic, dominated by high-discharge events associated with precipitation from the passage of hurricanes and extratropical storms (Hill *et al.* 2004), and by annual spring floods (freshets) governed by melting snow and enhanced rainfall (Brothers *et al*. 2008). Although these high-discharge events feed sediment to barrier systems across the GoM (FitzGerald *et al.* 2002), they are especially important at the mouth of the Saco

River, where nearly all sediment is trapped in the estuary during normal flow conditions (Brothers *et al.* 2008).

Modern sediment redistribution by marine and coastal processes

Estuarine sediment trapping, salt-marsh development and barrier dynamics continue to the present day. Offshore, the presence of abundant active bedforms along the shallow shelves off the Merrimack and Kennebec rivers indicates that paraglacial deposits undergo varying degrees of reworking by waves, tides and currents (Dickson 1999; Hein *et al.* 2007). Prevailing summer wind throughout the GoM is from the south-southeast and produces low-energy wave conditions and swells (Bigelow 1924; Jensen 1983). Spring, fall and winter prevailing winds are from the west-northwest, and storm events are associated with the passage of high-pressure fronts from the northwest, inland low-pressure systems and northeast storms that parallel the coast (Hill *et al.* 2004). Nor'easters (macrostorms driven by northeastern wind) account for at least 50% of all winter storms (Dolan & Davis 1992) and produce the strongest wind and waves across the GoM. Along the mixed-energy, east- to northeast-facing, drift-aligned coastline at the mouth of the Merrimack River, these nor'easters drive southerly longshore transport at a rate of 38,000 (Castle Neck) to 150,000 (Plum Island) m^3/yr (Smith 1991). By contrast, Saco Bay is more sheltered; here, wave refraction around headlands leads to a net northeast longshore transport estimated at only 10,000–16,000 m^3/yr (Kelley *et al.* 2005). The southward-facing coast near the mouth of the Kennebec/Androscoggin river system is largely protected from northeast waves by bedrock headlands. This swash-aligned system experiences a clockwise sediment gyre driven by tidal currents and storm waves (FitzGerald *et al.* 2000).

Tides in the GoM are semidiurnal and ranges generally increase from 2.5 m at Cape Cod in the southwest to *ca.* 17 m in the Bay of Fundy in the northeast. The shorelines at the mouths of the Kennebec/Androscoggin, Saco and Merrimack rivers all experience similar spring tidal ranges (2.9–3.1 m), though tidal prisms vary markedly among them as a function of the area and nature of backbarrier environments (percentage covered by tidal flats, bays and/or marshes), river dimensions and anthropogenic modifications (Table 1) (FitzGerald *et al.* 2005). Circulation in the present GoM, which has a mean depth of ~140 m, is generally cyclonic and dominated by buoyancy-driven coastal currents (Beardsley *et al.* 1997; Lynch *et al.* 1997; Lentz 2012) that have little impact on the coast except during major storms.

River-associated paraglacial barrier development: Examples from three barrier complexes

The combination of variable structural controls, sediment sources and supply rates, RSL changes and hydrographic regimes has produced a diverse set of barrier systems in the GoM. Three of the best studied are the Kennebec barrier chain (Kennebec and Androscoggin rivers), the Saco Bay barrier system (Saco River) and the Merrimack Embayment barrier chain (Merrimack River) (Fig. 1; Table 1). Each has distinctive features and a middle- to late-Holocene history of barrier formation and development. Evolutionary models for these systems are used here to contrast barrier formation along river-associated paraglacial coasts in different settings of sediment supply and accommodation, and exhibiting diverse intra-system variability.

Kennebec barrier chain

Located at the mouth of the Kennebec/Androscoggin river system, the Kennebec barrier chain (KBC) is located along a fjard-type paraglacial coast; that is, one formed along flooded glacial valleys with moderately shallow depths and moderate relief. The KBC consists of approximately 30 coastal accumulation forms (welded barriers and mainland beaches) stretching between New Meadows Bay (west of Small Point Harbor) and Sheepscot Bay (Fig. 6). These are subdivided into four physiographic provinces (West, Central, East-Central and Eastern complexes) that span nearly 250º in shoreline orientation (Fig. 6).

The KBC is fed by the Kennebec and Androscoggin rivers, which join at Merrymeeting Bay about 20 km north of the estuary mouth, and continue toward the GoM in a narrow bedrock-carved channel (Figs 3a & 6). The confluent, lower Kennebec River is a partially mixed to stratified mesotidal paraglacial estuary with seasonal variations in river discharge (Fenster & FitzGerald 1996). This river system continues to supply coarse-grained sediment to the coastal region, especially during spring freshets (Fenster & FitzGerald 1996). Most of this sediment is derived from upland outwash deposits, with compositions inherited from distinct bedrock lithologies (Borns & Hagar 1965). Today, the Kennebec River Estuary seaward of the Merrymeeting Bay receives a mixture of sediments from both the Kennebec and Androscoggin rivers; their sources can be differentiated on the basis of contrasting mineralogies that reflect the compositional differences of the respective river drainage basins (FitzGerald *et al*. 2002).

The KBC exemplifies an indented paraglacial coast that has experienced active but localized riverine sediment contribution to Holocene accumulation forms. All incipient transgressive barriers proximal to the Kennebec River mouth were established approximately 4.6 ka (Buynevich 2001). Away from the direct river influence, along sediment-starved complexes supplied primarily from the abandoned early Holocene deposits of the Kennebec River palaeodelta, transgressive barriers did not form until 1.2 ka (Buynevich 2001). The proximity to fluvial and palaeodeltaic sediment reservoirs and changes in three-dimensional accommodation space during decelerating RSL rise have been the major factors controlling the timing of barrier emplacement and progradation (Buynevich & FitzGerald 2001, 2005), degree of compartmentalization, and sediment volume (Barnhardt *et al*. 1997). Thus, despite their proximity and similarities, each complex of the KBC has a unique evolutionary history.

Situated on the western margin of the Kennebec estuary mouth, the Popham Beach System consists of a 4-km-long sandy barrier subdivided into three segments (Riverside, Hunnewell, and Seawall beaches) (Fig. 7a) anchored to pegmatitic bedrock headlands and isoclinally folded metasedimentary formations. Intertidal shoals connect the western and eastern ends of Hunnewell Beach to Fox Island and Wood Island, respectively (Figs 6 & 7). Extensive geophysical (ground-penetrating-radar (GPR)) and sediment-core data show that the Popham System contains at least 14 x 10^6 m^3 of Holocene sand (Buynevich & FitzGerald 2005). A texturally and compositionally submature, fluvially derived transgressive unit at -1.5 to -7.0 m MSL is confined to the landward part of this system, and can be traced alongshore for over 4 km, forming the core of the Hunnewell and Seawall beaches. Changes in shoreline exposure and decreasing proximity to the estuary mouth as the transgression proceeded led to an increase in maturity of subsequently deposited barrier sediments. Minimally reworked facies of fluvial origin were covered by younger parasequences composed of relatively mature, micaceous sand with evidence of estuarine sorting (Riverside Beach).

East of the Kennebec River Estuary, the Eastern Complex of Reid State Park (Fig. 6) demonstrates the effect of structural controls and relict offshore sediment sources on its Holocene development. This complex consists of two segments separated by a bedrock ridge:

Mile Barrier to the east and Half Mile Barrier to the west (Fig. 6). Mile Barrier is backed by a salt-marsh along its entire length, extending up to 600 m into the bedrock re-entrants. A small, man-made bedrock-bound inlet provides exchange between the ocean and the backbarrier. Numerous ledges and islands represent offshore structural extensions of bedrock ridges. These structural extensions can be followed for more than 10 km offshore in seismic-reflection profiles (Fig. 8) (Belknap *et al*. 1989; Buynevich *et al*. 1999), where they likely served as anchor points and compartmentalisation elements for early barriers. The coarse to very coarse-grained sand, particularly along the Mile Barrier, forms a steep (reflective) beach backed by a narrow dune ridge. GPR profiles and vibracores show an extensive salt-marsh unit beneath the barrier lithosome on either side of Todd's Head promontory that separates Mile and Half Mile Beaches (Buynevich & FitzGerald 2002). The presence, extent and age of this unit indicate that the salt-marsh functioned as a supratidal backbarrier platform for overwash and aggradation as early as 3000 years ago (Buynevich 2001).

The regressive/aggradational phases of barrier development along the KBC were marked by punctuated barrier progradation and dune development (Buynevich & FitzGerald 2001; Buynevich *et al*. 2004). Barrier progradation phases, which lasted for tens to hundreds of years, are recorded in the stratigraphy of the Popham Beach System as uniform to complex progradational units truncated by erosional scarps that are marked by high concentrations of heavy minerals (Fig. 7c). By contrast, along many of the small, sand-starved systems of the Western and East-Central complexes (Fig. 6), regressive barrier elements range from a single beach or dune ridge to widespread wind-driven aggradational units. Formation of these barriers, including some exhibiting limited progradation, was likely stimulated by a combination of a deceleration in RSL rise and increasingly efficient alongshore sand delivery facilitated by sediment filling of re-entrants (Buynevich 2001). Localized areas of retrogradation (Buynevich & FitzGerald 2005) reflect intense seaward-side barrier erosion as indicated by extensive dune scarps and heavy-mineral concentrations caused by storm surges and/or tidal-inlet migration (Fig. 7c) (Buynevich *et al.* 2004).

Depending on sediment supply and wind direction (for example, the large south-facing beaches of the Central Complex vs. the small north-facing pocket beaches within the Western Complex), internal sediment reworking and landward transport have resulted in growth and limited migration of parabolic and transgressive (climbing) dunes (Buynevich & FitzGerald 2003). The dune facies that presently comprise up to 40–50% of the barrier lithosome represents no more than 5–15% (~200 years) of the barrier history.

Saco Bay barrier system

Two prominent headlands composed of Paleozoic metamorphic rocks (Osberg *et al.* 1985), Prouts Neck and Biddeford Pool, delimit the 15-km-long Saco Bay barrier system (Fig. 9). This system is composed of relatively narrow barriers that are compartmentalized by bedrock ridges and pinnacles, tidal inlets, and, in the south, the Saco River Estuary. Differential erosion of Palaeozoic basement rocks by Paleogene and Neogene fluvial processes and Quaternary glacial processes has resulted in irregular topography that has given the coastal area its islands, headlands, and embayments.

The Saco River, which presently contributes 10,000 to 16,000 m^3/yr of sand to Saco Bay (Kelley *et al.* 1995c), primarily during spring freshets (Brothers *et al.* 2008), has supplied much of the sediment to the Saco Bay barrier system over the course of the Holocene (Kelley *et al.*

1995c; FitzGerald *et al.* 2005). During deglaciation, large amounts of glaciofluvial sand and gravel were deposited between the ice sheet and the ocean at many locations in the Saco River drainage basin and on the modern shallow shelf. These easily erodible sediments have been a particularly important source for the Saco Bay dune, beach, and shoreface systems (D. Barber, personal communication). Additional sediment has come from the erosion of glacial deposits near Prouts Neck (Kelley *et al.* 2005c).

GPR and core data allow the distinction of one longshore and three cross-shore barrier-sequence types within the Saco Bay barrier system (Fig. 10) (Van Heteren 1996). Highly diverse morphostratigraphies are reflective of several phases of middle- to late-Holocene barrier development. The Saco Bay barrier system shows a strong imprint of overstepping, fostered by the presence of numerous anchor points in the form of bedrock pinnacles, Pleistocene till mounds and other highs (Fig. 11) (van Heteren 1996). Backbarrier peat extends beneath only limited parts of the present-day barrier system, primarily on its landward margin, next to the modern salt-marshes. Inorganic backbarrier facies are much more widespread, underlying much of the modern barrier lithosome (Fig. 10) and occurring in the subsurface of the present-day inner shelf. This evidence of an ancient, wide, open-water lagoon or estuary extending well beyond the location of the modern barrier system indicates that an early barrier system formed seaward of the modern barriers during the early to middle Holocene (van Heteren 1996), possibly coincident with the sea-level slowstand of *ca.* 11.5 to 7.5 ka.

The precursor barriers deteriorated between 7.0 and 4.5 ka. This likely occurred in conjunction with the initial establishment, and subsequent longshore accretion, of proto-barriers that formed in the approximate location of the modern barriers (van Heteren 1996). This barrier morphosome initially abutted the mainland, at the edge of the former lagoon. In time a narrow backbarrier region developed through submergence of upland induced by rising RSL, allowing salt-marsh development and expansion (van Heteren 1996). Limited burial of backbarrier salt-marsh as a result of slow, RSL-induced barrier retrogradation has not compensated salt-marsh encroachment onto the mainland; this has resulted in the gradual widening of the backbarrier areas (van Heteren 1996).

The barrier system underwent limited retrogradation and continued longshore accretion until about 1 ka. Tidal inlets narrowed and the barrier system grew more continuous during this time. Finally, in the last 1000 years, most inlets closed or narrowed substantially to their modern dimensions and the barrier has primarily prograded seaward (van Heteren 1996).

The modern Saco Bay coastline has a distinct log-spiral (or zeta [ζ]) shape, reflecting the refraction pattern of oblique incident waves (Farrell 1970). Minor deviations from this log-spiral form are related to the presence of islands and submerged bathymetric highs off the present-day coast (Bremner & LeBlond 1974). The subaerial beach and dune system contains approximately 2.2×10^7 m^3 of sand (Kelley *et al.* 2005c). Relatively narrow, segmented barriers are exposed to waves from the easterly quadrant, driven by the region's dominant winds. Breaker heights are $<$ 0.5 m for the majority of the year, but during storms they average between 0.9 and 1.4 m (Farrell 1972). Net northerly transport along the beaches, at an estimated rate of 17,000 m^3/yr, results in long-term infilling of Scarborough River Inlet, a sediment trap located at the northern end of the barrier system (Kelley *et al.* 2005c). In the south, construction of jetties at the mouth of the Saco River in the mid-1800s has greatly impacted the adjacent shoreline and may have resulted in closure of the former Little River Inlet and rapid progradation of Pine Point (Fig. 9). This human measure resulted in the evolution of a secondary sediment sink between the Saco River jetties, which has necessitated regular dredging of the inlet at a rate of 10,000 m^3/yr (Kelley *et al.* 2005c).

Merrimack Embayment barrier system

The longest barrier chain in the GoM is located in the Merrimack Embayment (Figs. 2, 12). This mixed-energy inlet-segmented (FitzGerald & van Heteren 1999) embayment contains a 34-km-long series of barriers, tidal inlets, estuaries and backbarrier sand flats, channels and marshes. Individual barriers are 2–13 km long, generally less than 1 km wide and are backed primarily by marsh and tidal creeks that typically expand to small bays near inlets (Smith & FitzGerald 1994). They are commonly pinned to bedrock or shallow glacial deposits (Fig. 13a). Each contains abundant, vegetated parabolic dunes that reach as much as 20 m in elevation. These are best developed along central and southern Plum Island and Castle Neck, reflecting abundant quartz sand associated with high rates of longshore transport.

Sediments for this system were dominantly derived from the 180-km-long Merrimack River. Like the Kennebec, Androscoggin and Saco rivers, the Merrimack is largely bedrock-controlled. Its headwaters are in the pluton-dominated White Mountains of New Hampshire (Fig. 3b), ensuring a steady supply of quartz-rich, sandy sediments to the embayment. The Merrimack River has delivered an average annual bedload volume of 4.16×10^4 m^3/yr since at least the mid-1900s (Hein *et al.* 2012). Assuming a stable flux over time, the volume of coarse sand and gravel delivered by the Merrimack since barrier pinning at 4 ka (166×10^6 m^3) can account for the entire volume of the barriers and tidal deltas of the Merrimack Embayment (*ca.* 137×10^6 m^3) (Hein 2012). These barrier / tidal delta sand volumes are dwarfed by the volume of finer, sandy estuarine sediment in the backbarriers of the Merrimack barrier chain (*ca.* 850 x 10^6 m^3) (Hein *et al.* 2011a), comparable in volume to the lowstand palaeodelta (1300 x 10^6 m^3) (Oldale *et al.* 1983). Additional Holocene sandy sediments deposited across the shallow shelf as a sand sheet total ~650 x 10^6 m^3 (Barnhardt *et al.* 2009; Hein 2012). Assuming modern fluxes, sandy sediment supply from the Merrimack River since the 14 ka lowstand would amount to no more than 500 x 10^6 m^3, only ~1/3 of the combined volume of the sand sheet and backbarriers, both of which postdate the lowstand. Thus, fluvial sediment-supply rates during the early Holocene were likely several times higher than at present. Additional sediment for the present barrier, backbarrier and/or shelf sand sheet was also contributed from the erosion of regressive and lowstand deposits; seismic-reflection profiles across the lowstand delta demonstrate the presence of a smooth, gently dipping erosional surface that truncates the upper parts of delta foresets (Fig. 13b), indicating complete removal of thin topset beds and scouring to an unknown depth during the early transgression (Oldale *et al.* 1983; Barnhardt *et al.* 2009). The relative contributions from the river and from marine reworking of Upper Pleistocene coastal and deltaic deposits are unknown.

The large supply of fluvial, glacial, and paraglacial sediment available to the Merrimack Embayment barriers has played a dominant role in their formation and subsequent development. Sediments provided by the erosion of the lowstand palaeodelta and the regressive braidplain delta during the period of relatively rapid RSL rise and shoreline transgression (*ca.* 12–6 ka), as well as the direct contribution by the Merrimack River, triggered the development of overstepping barriers and transgressive sand shoals (FitzGerald *et al.* 1994). Thin and mobile protobarriers were pinned to contemporaneous emerged drumlins and bedrock outcrops proximal to modern barrier positions by 4 ka (Figs 13a & 14a) (Hein *et al.* 2012). There is little evidence of overwash following this original pinning phase. Abundant GPR profiles collected along Salisbury Beach and Plum Island reveal few overwash deposits along the proximal landward side of the barriers (Fig. 15a, b) (Costas & FitzGerald 2011; Hein *et al.* 2012). Although preserved washovers are somewhat more prevalent along the Crane Beach section of Castle Neck (Dougherty *et al.* 2004), southerly and seaward-dipping reflections generally dominate (greater than 90%) barrier widths (Fig. 15c). This is indicative of the several thousand years of progradation, aggradation and spit elongation that built the modern barriers (Hein *et al.* 2012).

Palaeo-inlet sequences are common along the Merrimack Embayment barrier chain (Fig. 14) (Hein *et al.* 2012). The closure of the palaeo-Parker Inlet, located in central Plum Island, highlighted the role of the deposition of abundant fine- to medium-grained estuarine sand in barrier stabilisation. Here, decelerating RSL rise at about 6 ka led to reduced mainland shoreline transgression and diminished creation of backbarrier accommodation. Backbarrier sedimentation exceeded accommodation creation over several thousand years, leading to a decrease in tidal prism and ultimately to inlet closure between 3.6 and 3.0 ka. This closure, which created a single 13-km-long island, was closely followed by a rapid expansion of backbarrier marshes and by the aggradation, elongation, and progradation of the barrier itself (Hein *et al.* 2011a, 2012).

Modern sediment supplied by the Merrimack River is largely driven southward along Plum Island towards Castle Neck by the dominant northeast storm waves. Southerly oriented, ebb-dominant sandwaves within the large ebb-tidal-delta complex corroborate sedimentological evidence of a southerly fining trend across the ebb delta, and of a general trend of increasing textural and mineralogical maturity in the same direction, away from the river (FitzGerald *et al.* 1994, 2002). The dominant southerly transport regime has resulted in the growth of recurved spits on the downdrift ends of Crane Beach and Plum Island (Farrell 1969), is reflected in an increase in the spacing of offshore contours to the south along the barrier chain (Smith 1991), and has influenced the development of the southern part of the Merrimack Embayment as a net sediment sink (Hubbard 1976; Barnhardt *et al.* 2009). Active transport across the shallow (less than *ca.* 40 m) shelf during northeast storms likely serves to rework some of this temporarily stored sediment toward the barrier system (Hein *et al.* 2007). Proximal to the mouth of the Merrimack River, continued fluvial sediment inputs combine with complex river / inlet / tidal-delta interactions to create the most dynamic section of this coast (Fig. 15a) (FitzGerald *et al.* 1994; Costas & FitzGerald 2011).

Paraglacial coasts: Barrier formation in a distinctive setting

River-associated barriers of the Gulf of Maine in a paraglacial context

Coastal barriers within a paraglacial coastal framework occur in a variety of settings, including braided outwash plains, estuarine re-entrant coasts, and prograding deltaic systems. These systems have responded to different RSL histories, sediment supplies, and physical processes of sediment reworking, but they have all built barriers, tidal inlets and associated tidal sand bodies, and various types of backbarrier environments. Similarities among these systems and differences within individual systems emphasize the importance of wave and tidal processes and RSL changes in dictating coastal morphology.

Closest to active glaciers, most barriers formed at the leading edge of exposed, prograding outwash plains, a setting found along the active Skeiðarársandur coast of southeast Iceland (Hine & Boothroyd 1978; Nummedal *et al.* 1987), the Gulf of Alaska coast from Dry

Bay west to Kayak Island (Hayes & Ruby 1994), the Hallo Bay region along the Alaskan Peninsula and the Karaginsky Gulf coast of the Kamchatka Peninsula, Siberia. Coastal accumulation forms of this type are marked by a wide range in grain sizes, but coarse sand and fine gravel dominate most barrier and tidal-delta lithosomes. These arctic and subarctic coasts are dominated by spit systems; true barrier islands are rare. Tidal inlets tend to be located at the downdrift end of littoral cells. Barriers fronting active, prograding outwash plains are generally susceptible to breaching during storms, a process that commonly repositions inlets in the middle of embayments. Flood-tidal deltas along these outwash-plain barriers tend to be well developed and are commonly a product of storm deposition, whereas ebb deltas are only prominent at inlets with large tidal prisms.

Farther from active ice margins, the largest paraglacial barriers are associated with fluvial systems draining glaciated landscapes. These have voluminous sand and gravel sources which are generally replenished less frequently as glaciers recede. Still within the sphere of influence of ice caps or glaciers, some systems are nourished by sediment eroded locally from outwash plains that are no longer fully replenished by meltwater-derived sediment, especially in the coarsest fractions, whereas others are supplied by major transport conduits from sediment supplied from farther away. These latter systems are found at the mouths of rivers draining active mountain glaciers (for example, the Copper River, Alaska; Hayes & Ruby 1994). They are commonly characterized by well-developed barrier islands and tidal inlets and by backbarriers with open-water (lagoonal) or intertidal (marsh, tidal flats and tidal creeks) dominance, dependent in part on antecedent setting. Sand dominates these systems because of the distal location of the primary sediment source, although fine gravel may also be an important component. The spacing and dimensions of inlets along these coasts are a function of tidal prism, whereas inlet positions and sizes on barrier coasts such as those fronting the Kennebec/Androscoggin, Saco and Merrimack rivers in the GoM are usually a function of flood events, basement controls, storm breaching, and prominent river-discharge sites.

Beyond the area of direct glacially driven sediment replenishment, most paraglacial barriers are formed and modified using finite and shrinking sand and gravel sources, both on- and offshore. Where these sources are large and exposed enough to supply transporting conduits and processes with a continuous and steady flow of sediment, the absence of replenishment does not influence barrier development (such as the Kennebec, Saco and Merrimack barrier chains). Where sources are small and localized, they will eventually become rapidly exhausted (for example, along the drumlin-dominated Eastern Shore of Nova Scotia; Carter *et al.* 1990; Forbes *et al.* 1991).

When viewing the GoM barriers in light of these proximal and distal settings, behavioural differences and temporal developmental patterns related to sediment availability can be explained. The most voluminous sediment resources along the GoM are located farthest south, near the terminus of the Wisconsinan ice sheet at Cape Cod. Here, early barrier systems will have been dominated by coarse-clastic spits. Even today, sediment eroded from exposed outwash bluffs nourishes attached barriers, some of which are periodically breached to become islands. On the other side of the spectrum, barrier systems beyond the area of continuing glacial sediment replenishment, and thus dependent on local sediment supplies, are most strongly affected by the finite nature of sediment sources. Examples include the barriers of the Eastern Shore of Nova Scotia and the barriers and spits of northwest Alaska. Similar to the river-associated GoM barriers, these barriers are generally composed of medium to coarse sand and gravel (Short 1979; Boyd *et al.* 1987); where present, backbarrier and estuarine sediments are composed of fine sand

and mud (Boyd & Honig 1992; Carter *et al.* 1992). However, sediment supply to river-associated barriers along the central Maine coast is much less variable. With rare exceptions, such as near the southern end of Plum Island where drumlins have been eroded episodically, middle- and late-Holocene sediment supply has been marked by a steady decrease (D. Barber, personal communication). Here, RSL changes have maintained a dominant differential control over barrier evolution during the past *ca.* 8000 years. The present organization of the Kennebec, Saco Bay and Merrimack Embayment barrier systems indicates an increasingly high degree of maturity that developed during an extended period (3000–5000 years) of lateral barrier accretion and barrier progradation under conditions of near-steady fluvial sediment supply to the coast. Earlier in their development, however, each of these systems may have shown a much higher morphologic diversity as they went through one or more stages of morphologic immaturity and deterioration during periods of rapid RSL rise and barrier retrogradation that dominated over sediment availability. The coexistence of immature, mature, and disintegrating barriers along these and other paraglacial coasts (Kliewe & Janke 1991; Orford *et al.* 1991; Nichol & Boyd 1993) is a morphological reflection of recent diachronicity in barrier development. Morphologic evidence from past developmental phases, however, is easily overprinted by later events.

Features of barrier formation in paraglacial settings

The paraglacial barriers of the GoM differ from well-researched lower-latitude barriers that are located far beyond the limits of the Pleistocene ice sheets in terms of both their development and morphostratigraphy (Table 4). Their development has been affected by: (1) numerous bedrock promontories that compartmentalized the coast and served as pinning points for barriers; (2)

spatially and temporally variable sediment sources (plutons, reworking of glacial and paraglacial sediments) and a wide range in particle sizes (clay to boulders) and sediment supply rates; ; and (3) multiple phases of transgression and regression, associated with rapid, large-scale, post-glacial sea-level changes (for example, RSL fall of 120 m over 2000 years in central Maine; Fig. 4) that allowed the reworking of sediment across the modern coastal plain, coastal zone and shelf. The resulting barrier systems are characterized by progradational dunes and beaches composed of spatially variable sediment textures; for example, beaches are coarser proximal to glacial (till) outcrops and fluvial sediment sources and fine with distance. They have complex mineralogical compositions, such as the presence of horizons rich in garnet derived from erosion of plutons, and an overall lithological heterogeneity such as glaciomarine silt and clay directly underlying coarse sandy barriers. Bioclastics within the barrier lithosomes are mid/late Holocene to modern in age. Preserved marine organisms are largely similar within GoM coastal sediments, with differences reflecting the presence of exposed bedrock. For example, barnacle plates are much more common in the bedrock-dominated KBC than in the glacially pinned Merrimack barriers.

Paraglacial GoM barriers are more commonly composed of coarser-grained sediment (siliciclastic medium-grained sand to gravel) than is found along non-glaciated continental trailing-edge coasts. This reflects their (1) proximity to major river systems with coarse sediment loads; (2) glacial sediment sources, and (3) continued inputs from both glacial (drumlins and other till deposits) and paraglacial (upstream terraces, outwash deposits) sediment sources.

Variable sediment supply: The defining feature of paraglacial barrier formation

Glacial and paraglacial settings are, by definition, transient: their existence and persistence are inextricably linked to the availability and supply of glacially derived sediment. The surficial evidence of glaciation along river-associated paraglacial coasts may be removed within several thousand years by non-glacial processes. Nonetheless, a glacial imprint remains, commonly hidden within the stratigraphic framework of fluvial and coastal deposits. The sedimentological, stratigraphic and chronologic frameworks of the river-associated barriers of the GoM contain valuable information on the nature and timescales of coastal landscape responses to glaciation.

The paraglacial period, as defined by Ballantyne (2002a, b), is the timescale over which glacigenic sediment stores are tapped and finally exhausted, or landscapes equilibrate to reworking processes. Following this phase, the landscape returns to a non-glacial or post-glacial state. As such, a coastline is only paraglacial for as long as glacially excavated landforms and glacigenic sediments have a recognizable influence on the character and evolution of the coast (Forbes & Syvitski 1994). The duration of the paraglacial period along any given coastline can be affected by regional geology (erodibility of bedrock) and by temporally varying sediment supplies. Sediment-supply variations can be driven by changes in climate (amount / seasonality of precipitation), vegetation (stabilisation of hill slopes), proximity to meltwater (steady meltwater fluxes; annual melting events; convulsive outburst and catastrophic flooding events (jökulhlaups)), regional coastal setting (wave; tides; river inputs) and RSL changes (for example, erosion of untapped glacigenic sediment sources by rising RSL) (Forbes & Taylor 1987; Forbes & Syvitski 1994; Forbes *et al.* 1995). Continued fluvial input of sediment eroded from voluminous glacial sources and complex post-glacial RSL changes along the river-associated paraglacial coast of the GoM served to lengthen the paraglacial period through the late Holocene and broaden its influence from inland to far offshore. The sediment yield of all river systems feeding these coasts is dominated by reworked glacial and paraglacial deposits. These serve only as temporary storage for sand and gravel generated thousands of years earlier. Elevated sediment export from river catchments will continue as long as glacigenic or paraglacial sediments remain easily accessible to fluvial scouring (Church & Ryder 1972; Ballantyne 2002a). The re-entrainment of submerged paraglacial deposits within the coastal setting is another mechanism lengthening the period of influence of glaciation. In the GoM, the reworking of regressive and lowstand deposits offshore of river mouths has produced transgressive sand sheets and nourished barrier systems. Ballantyne (2002a, b) suggested the term “secondary paraglacial system” to describe such features. In this case, the regressive and lowstand deposits can be considered primary paraglacial deposits and the later barriers secondary paraglacial landforms.

The delivery of sediment to river-associated coastal paraglacial systems during and after glaciation occurs over multiple periods, each marked by the deposition of distinct sedimentary units and the formation and modification of specific landforms (Figs 5 & 16):

Glacial period

Characteristic deposits and landforms during this period are formed by glaciers during both advance and retreat of ice sheets. These features include glacially striated bedrock, fjords, over-steepened embayments, drumlins, crag-and-tail ice-streamlined deposits, kames, eskers, grounding-line fans, ground and washboard moraines and deglaciation-related moraines (lateral, terminal, recessional; submarine ice-pushed moraines) (Belknap *et al.* 1987; Syvitski 1991; Forbes & Syvitski 1994). Till is the dominant deposit.

Proglacial period

The proglacial period begins immediately following deglaciation (Church & Ryder 1972; Ballantyne 2002a,b; Slaymaker 2009). Although the term *proglacial* specifically refers to an area literally "in front of the glacier", it is adopted here for a time period during which a landscape is located in such a position because of retreat of an ice sheet. Sediments deposited during this period are largely derived directly from the glacier and are therefore *glacial* rather than *paraglacial* in origin.

In the Gulf of Maine, the proglacial period was marked by deep isostatic depression of the crust, rapid shoreline transgression, the presence of tidewater glaciers and the deposition of outwash deposits and coarse sandy ice-marginal deltas. Elsewhere, proglacial coastal-zone sediments accumulate in sandur (outwash) plains, braided outwash fans, jökulhlaup units, glaciomarine deltas and coastal moraines, ice-rafted debris deposits, and glaciomarine basins (Syvitski 1991; Slaymaker 2009). Along river-associated coasts in the GoM, fluvial sediment delivery to the coast during this period was characterized by finer mean grain sizes than during the glacial period (Fig. 5). The proglacial period can last for many (> 10) millennia if active glaciers continue to indirectly deliver proglacial sediments to the coast via wide braidplains (for example, Skeiðarársandur, Iceland).

Early paraglacial period

This period is newly defined here for river-associated paraglacial coasts, differentiating them from other paraglacial coastal settings. The onset of this period coincides with the end of widespread glaciation within the barrier-associated river drainage basin. Glacigenic sediments are no longer being deposited within the drainage basin or along the coast. Sand-sized sediment export to the coast reaches a maximum: a dearth of vegetation and soil, and large quantities of glacially liberated and paraglacial sediment stored within the river basins at the start of this period, lead to a high rate of sediment - notably sand - export. High fluvial sediment yields, obviously proportional to river discharge, continue throughout this period. Along the coast, this early paraglacial period may be referred to as the ***paraglacial sand maximum***, a time of unusually high rates of sand delivery to, and deposition in, various contemporaneous coastal environments. Sediments supplied by rivers are reworked by coastal and fluvial processes in a regressive setting, in response to isostatic-rebound-induced RSL fall. In the GoM, this mechanism resulted in the deposition of both discrete and expansive (for example, the Sanford-Kennebunk, Brunswick and Merrimack braidplain deltas) regressive shoreline deposits and lowstand deltas and delta lobes. During this time of rapid isostatic land-level adjustments in the drainage basins of paraglacial rivers, drainage divides of large and small streams shift and can strongly alter both the competence of streams and the distribution river mouths.

Middle paraglacial period

This period of coastal paraglacial evolution is marked by a gradual exhaustion of terrestrial glacigenic and paraglacial deposits by associated river systems and by a commensurate decrease in fluvial sediment supply to the coast. Intense reworking by waves, tides and currents of glacigenic landforms (drumlins, till bluffs, onshore outwash plains) and newly drowned primary coastal paraglacial deposits (regressive shoreline deposits, lowstand deltas) results in the formation of secondary coastal paraglacial deposits and features such as transgressive sand sheets, estuarine fills, barriers, tidal deltas, backbarrier marshes and tidal flats. Rising RSL eventually fully submerges and strands glacigenic and primary paraglacial deposits beyond the depth of reworking by waves as transgression progresses. However, this same RSL rise and shoreline transgression can also tap new sediment sources and alter the distribution and prominence of pinning points and coastline orientation, resulting in cyclic patterns of barrier formation and destruction such as found along drumlin-dominated coasts (Rosen 1984; Forbes & Taylor 1987; Nichol & Boyd 1993). A key feature of this period along river-associated barrier coasts is the transition to dominantly non-glacigenic and non-paraglacial fluvial sediment export. This reflects the development of post-glacial conditions upstream, in which sediment transport and export is in equilibrium with erosion of primary materials (bedrock) by non-glacial processes. Such a transition may take as long as 5000 to 10,000 years as even bedrock erosion rates may remain elevated for long periods because of the persistence of a weakened upper bedrock surface that had been physically fractured and weathered by a long-departed ice sheet.

Late paraglacial period

This stage is marked by the gradual exhaustion of accessible paraglacial sediment. During this period, coastal systems are fed dominantly by rivers that export sediment derived only from erosion of non-glacial sources, such as the bedrock in New England. Non-fluvial sediment inputs to the coastal system are increasingly rare. Some final paraglacial activity is linked to occasional erosion of coastal glacigenic or paraglacial deposits, but most of the non-fluvial sediment will be derived from shorefaces and shelves that are increasingly in equilibrium with prevailing wave and current conditions. Along sediment-rich coasts such as at the mouths of the Kennebec/Androscoggin, Saco and Merrimack rivers in the GoM, deposits associated with this period largely reflect the maturation of coastal features first formed during the middle paraglacial period. By contrast, along sediment-starved coasts or those with a high rate of creation of accommodation space for a given RSL rise (that is, those with a low gradient), the period of transition to negligible (or intermittent) glacigenic / paraglacial sediment supply can lead to widespread erosion, barrier instability and eventual disintegration.

In the GoM, the transition to the late paraglacial period corresponded with a marked decrease in the rate of RSL rise at *ca.* 4 ka (Fig. 4). Relatively mature fluvial and reworked paraglacial sediments filled an increasing proportion of available accommodation along river-associated coasts in the GoM. Progradation, aggradation, and the development of an equilibrium shoreface, expansive vegetated and unvegetated dunes, marshes, and equilibrium ebb and flood-tidal deltas were the result.

Post-paraglacial period

This period is characterized by a coast that is controlled exclusively by non-glacial processes and does not receive any additional sediment from any glacigenic or paraglacial sources. The coast responds to intrinsic and extrinsic forcings in a manner indistinguishable from a coast in a non-glaciated setting.

The timing of the transition to, and even the existence of, the post-paraglacial period is highly contentious for all areas falling within the sphere of influence of Quaternary glaciations. Coastal systems within these areas are never fully and permanently removed from the effects of glaciation. After many years of non-paraglacial conditions, a storm may expose a glacial source, or a changing coastal or fluvial configuration may make one barrier paraglacial again and render another barrier non-paraglacial for the first time. For example, the river-associated paraglacial barriers in the GoM, although formed solely by non-glacial processes, would not have had a sediment source large enough for their formation without the prior deposition of paraglacial deposits. Even after sediments have been reworked multiple times, these retain imprints, either clear or very subtle, of their paraglacial origins. Furthermore, glacigenic and paraglacial sediments can be released in terrestrial environments by processes unrelated to direct or lagged landscape response to glaciation. For example, changes in climate or human disturbance can rapidly change terrestrial landscapes, affecting erosion and deposition patterns and delivering fresh quantities of previously unavailable glacigenic or paraglacial sediment to a river-associated paraglacial coast many millennia following deglaciation. Likewise, increasing rates of RSL rise, notably along the northeast coast of the USA (Sallenger *et al.* 2012) will force shorelines to transgress previously untapped terrestrial glacigenic sediment sources, thus contributing new glacigenic sediments to the barrier systems. In the Merrimack Embayment, drumlins presently undergoing erosion at the southern end of Plum Island and on Castle Neck are cored by till of Illinoian age (Stone *et al.* 2006); erosion of these would contribute glacigenic sediments that have been stored along the coast for more than 100,000 years. Does this imply that the Merrimack Embayment is still in a state of paraglacial or post-paraglacial non-equilibrium dating back to approximately 120 ka? A positive answer to this question implies that coasts in regions susceptible to glaciations may never truly enter a post-paraglacial period and will always be paraglacial in nature. Using this line of reasoning, even the major North Sea barrier system extending from northern France to Denmark is entirely paraglacial. Along the Frisian Islands, a clear paraglacial overprint is visible in eroding bluffs (Denmark and Germany) and in glacial highs serving as anchor points to barrier islands (Denmark, Germany and the Netherlands). Farther south, the paraglacial characteristics are more subtle. Here, the progradational barrier system of the western Netherlands owes much of its size to a mid-Holocene abundance of fluvial lowstand sand formed under periglacial conditions and later reworked by shallow-marine and coastal processes as the initially gentle shoreface steepened toward equilibrium.

Changing climate, human interference and the future of river-associated barriers in the Gulf of Maine

Coastal sediment supplies have undergone significant natural and human-induced perturbations at local, regional and global scales and over time periods ranging from months to thousands of years. Natural climate-geomorphic feedbacks under changing precipitation regimes have driven changes in the rates of erosion and fluvial sediment delivery to the coast (Leeder *et al.* 1998; Blum & Törnqvist 2000; Goodbred 2003; Hein *et al.* 2011b). Over the Anthropocene, fluvial sediment supplies have further varied in response to deforestation, agricultural expansion and contraction, urbanization, sediment quarrying and mining, land reclamation, and river engineering, impoundment and damming (Yang *et al.* 2010; Kirwan *et al.* 2011; Milliman & Farnsworth 2011). Over shorter timescales, the emplacement of artificial hard protective structures (such as jetties, groins, seawalls, bluff-stabilisation measures, and breakwaters) and implementation of soft engineering solutions (beach and shoreface nourishment, dewatering, sand-bagging, scraping and draining) have disrupted natural pathways of sediment within the littoral zone, resulting in migration of accretion and erosion hotspots, modification of overall beach morphodynamics, and localised flooding. Given projected increases in the rates of RSL rise (Church *et al.* 2014), net coastal erosion resulting from both human-induced and natural changes in sediment-supply systems will likely only accelerate.

Changing climate and enhanced anthropogenic stresses present unique challenges for the future viability of river-associated paraglacial barrier coasts in particular. In contrast to coasts proximal to actively melting glaciers that are fed by sediment-laden meltwater streams or sustained by sediment eroded from coastal bluffs composed of thick glacial deposits, those with thinner or less extensive glacigenic and paraglacial inland deposits are more likely to be impacted by a future reduction in terrestrial sediment supply. The GoM barriers are presently nourished by rivers with discharges similar to those of many moderate-sized rivers along the East and Gulf coasts of the USA (Milliman & Farnsworth 2011); however, natural depletion of glacial and paraglacial fluvial sediment sources and the stabilisation of slope, terrace and floodplain deposits by vegetation cover have reduced fluvial sediment supply to these paraglacial coasts more significantly than farther south. Unlike barriers in non-glaciated or river-distal settings, river-associated paraglacial barriers formed in regimes of both ample accommodation and ample fluvial sediment supply, but now face a substantial natural and anthropogenically induced reduction in the latter. The drainage basins, tributaries and primary downstream river segments of the Kennebec / Androscoggin, Saco and Merrimack rivers have all undergone extensive anthropogenic modifications over the past several hundred years. Damming and re-routing of the rivers and their tributaries and the jettying and dredging of their mouths have greatly impacted sediment discharge and sand-dispersal patterns (Farrell 1970; FitzGerald 1993; Kelley *et al.* 2005c).

Given the reductions in fluvial sediment supply and human alterations of fluvial sediment-supply systems, enhanced creation of backbarrier accommodation in a regime of accelerated RSL rise is likely to cause narrowing and shortening of the river-associated GoM barriers (FitzGerald *et al.* 2008), with an increasing chance of barrier breaching. This process may reach the point where some barriers become unstable and vulnerable to the step-wise retrogradation and overstepping that characterized their earlier histories. The retrogradational/aggradational pathway of the Kennebec barrier chain, for example, will likely continue in the coming decades of accelerated RSL rise (Buynevich 2001). As attested by occasional intertidal exposures of backbarrier sediments and tree stumps on the beach face, onshore-offshore redistribution of sand and gravel during intense storms will continue to drive longer-term barrier morphodynamics (Buynevich *et al* 2004). Mobility of the mature barrier-spit systems of Saco Bay and the Merrimack Embayment is likely to be limited in the near term. However, when sediment supply no longer compensates the effects of RSL rise, even these barrier spits will either migrate rapidly to a nearby pinning point, as they have in the past (van Heteren 1996; FitzGerald *et al.* 1994; Hein *et al.* 2012), or be fragmented and destroyed, only to reform in a more landward position that is favoured by palaeotopography (Swift 1968; Boyd & Penland 1984). Thus, river-associated paraglacial coasts dependent on continuous fluvial sediment input are in a precarious situation and may be rapidly approaching a point of transition from regressive to transgressive and destructive modes.

Conclusions

Paraglacial coasts are those formed on or proximal to formerly ice-covered terrain and retain the landforms and sediments derived directly or indirectly from glaciation. The river-associated paraglacial barriers of the GoM (USA & Canada) formed along such a paraglacial coast during a period of decelerating RSL rise over the past *ca.* 5000 years. These barriers are distinguishable from barriers formed in coastal plain settings or even those along other paraglacial coasts. They are characterized by spatially variable sediment textures, complex sediment composition and lithological heterogeneity. This variability reflects several unique features of barrier formation along river-associated paraglacial coasts: (1) the abundance of bedrock and glacial promontories that compartmentalize the coast and serve as pinning points for barriers; (2) the complex post-glacial RSL changes that can shift depocenters laterally tens of km in hundreds of years; and (3) the variable sources, conduits, and supply rates of glacial, primary and secondary paraglacial, and nonglacial sediment sources. Sediment-supply rates along these river-associated paraglacial barriers were highest within a few thousand years following deglaciation (the early and middle paraglacial periods). Sand deposition peaks at a period herein defined as the *paraglacial sand maximum*, as glacial and primary paraglacial deposits are eroded on land and sediments are redeposited along the regressing coast as a series of sandy shorelines, braidplain deltas and lowstand deltas.

The future stability of river-associated paraglacial barriers in a regime of accelerated RSL rise is dependent upon the continued supply of sandy sediments to the barriers and beaches, and of finer inorganic sediments to the backbarriers. However, a combination of a natural depletion of glacially liberated sediment and anthropogenic modifications of both the river systems delivering this sediment and the barriers themselves threatens to enhance barrier erosion, cause disintegration of backbarrier marshes, and, eventually, return these systems to the retrogradational states that characterized their earlier development.

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References

BALCO, G., STONE, J. O. H., PORTER, S.C. & CAFFEE, M.W. 2002. Cosmogenic-nuclide ages for New England coastal moraines, Martha's Vineyard and Cape Cod, Massachusetts, USA. *Quaternary Science Reviews*, **21**, 2127–2135.

BALLANTYNE, C. K. 2002a. Paraglacial geomorphology. *Quaternary Science Reviews*, **21**, 1935–2017.

BALLANTYNE, C. K. 2002b. A general model of paraglacial landscape response. *The Holocene*, **12**, 371–376.

BARNHARDT, W. A., GEHRELS, W. R., BELKNAP, D. F. & KELLEY, J. T. 1995. Late Quaternary relative sea-level change in the Western Gulf of Maine: Evidence for a migrating glacial forebulge. *Geology*, **23**, 317–320.

Barnhardt, W. A., Belknap, D. F. & Kelley, J. T. 1997. Stratigraphic evolution of the inner continental shelf in response to late Quaternary relative sea-level change, northwestern Gulf of Maine. *Geological Society of America Bulletin*, **109**, 612–630.

Barnhardt, W., Andrews, B., Ackerman, S., Baldwin, W. & Hein, C. 2009. *High Resolution Geologic Mapping of the Inner Continental Shelf: Cape Ann to Salisbury Beach Massachusetts*, United States Geological Survey Open-File Report **OFR–07–1373**.

Beardsley, R. C., Butman, B., Geyer, W. R. & Smith, P. 1997. Physical Oceanography of the Gulf of Maine. *In:* Wallace, G. T. & Braasch, E.F. (eds) *Proceedings of the Gulf of Maine Ecosystem Dynamics Scientific Symposium and Workshop*, Regional Association for Research on the Gulf of Maine RARGOM, **Report 97-1**, 39–52.

Belknap, D. F., Kelley, J. T. & Shipp, R. C. 1987. Quaternary stratigraphy of representative marine estuaries: Initial examination by high resolution seismic reflection profiling. *In:* FitzGerald, D. M. & Rosen, P.S. (eds) *Glaciated Coasts*. Academic Press, New York, 143–176.

Belknap, D. F., Shipp, R. C., Kelley, J. T. & Schnitker, D. 1989. Depositional sequence modeling of the late Quaternary geologic history, west-central Maine coast. *In:* Tucker, R. D. & Marvinney, R. G. (eds) *Studies in Maine Geology - Vol. 5: Quaternary Geology*. Maine Geological Survey, Augusta, 29–46.

Belknap, D.F., Kelley, J.T. & Gontz, A.M. 2002. Evolution of the glaciated shelf and coastline of the Northern Gulf of Maine, USA. *Journal of Coastal Research*, **SI 36**, p. 37–55.

Bigelow, H. B. 1924. Plankton of the offshore waters of the Gulf of Maine. *Bulletin of the U.S. Bureau of Fisheries*, **40**, pp. 1–509.

Bloom, A. L. 1963. Late-Pleistocene fluctuations of sea level and postglacial crustal rebound in coastal Maine. *American Journal of Science*, **261**, 826–879.

Blum, M. D. & Törnqvist, T. E. 2000. Fluvial responses to climate and sea-level change: a review and look forward. *Sedimentology*, **47**, 2–48.

Borns, H. W., & Hagar, D. J. 1965. Late-glacial stratigraphy of a northern part of the Kennebec River valley, western Maine. *Geological Society of America Bulletin*, **76**, 1233–1250.

Borns, H. W., Jr., Doner, L. A., Dorion, C. C., Jacobson, G. L., Jr., Kaplan, M. R., Kreutz, K. J., Lowell, T. V., Thompson, W. B. & Weddle, T. K. 2004. The deglaciation of Maine, USA. *Developments in Quaternary Sciences*, **2**, 89–109.

Boyd, R. & Penland, S. 1984. Shoreface translation and the Holocene stratigraphic record: Examples from Nova Scotia, the Mississippi Delta and eastern Australia. *In:* Greenwood, B. & Davis, R. A. Jr. (eds) Hydrodynamics and Sedimentation in Wave-Dominated Coastal Environments. *Marine Geology*, **60,** 391–412.

Boyd, R. & Honig, C. 1992. Estuarine sedimentation on the Eastern Shore of Nova Scotia. *Journal of Sedimentary Research*, **62**, 569–583.

Boyd, R., Bowen, A. J. & Hall, R. K.1987. An evolutionary model for transgressive sedimentation on the Eastern Shore of Nova Scotia. *In:* FitzGerald, D. M. & Rosen, P.S. (eds) *Glaciated Coasts*. Academic Press, New York, 87–114.

Bremner, J. M. & LeBlond, P. H. 1974. On the planimetric shape of Wreck Bay, Vancouver Island. *Journal of Sedimentary Petrology*, **44**, 1155–1165.

Broecker, W. S., Turekian, K. K. & Heezen, B. C. 1958. The relation of deep sea [Atlantic Ocean] sedimentation rates to variations in climate. *American Journal of Science*, **256**, 503–517.

Brothers, L. L., Belknap, D. F., Kelley, J. T., & Janzen, C. D. 2008. Sediment transport and dispersion in a cool-temperate estuary and embayment, Saco River estuary, Maine, USA. *Marine Geology*, **251**, 183–194.

Buynevich, I. V. 2001. Fluvial-marine interaction and Holocene evolution of sandy barriers along an indented paraglacial coastline. PhD thesis, Boston University.

Buynevich, I.V. & FitzGerald, D.M. 2001. Styles of coastal progradation revealed in subsurface records of paraglacial barriers, New England, USA. *Journal of Coastal Research,* **SI 34**, 194–208.

Buynevich, I. V**.** & FitzGerald, D. M. 2002. Organic-rich facies in paraglacial barrier lithosomes of northern New England: preservation and paleoenvironmental significance. *Journal of Coastal Research*, **SI 36**, 109–117.

Buynevich, I. V. & FitzGerald, D. M. 2003. High-resolution subsurface (GPR) profiling and sedimentology of coastal ponds, Maine, USA.: Implications for Holocene back-barrier evolution. *Journal of Sedimentary Research*, **73**, 559–571.

Buynevich, I. V. & FitzGerald, D. M. 2005. Sedimentological signatures of riverine-dominated phases in estuarine and barrier evolution along an embayed coastline. *In:* FitzGerald, D. M. & Knight, J. (eds) *High Resolution Morphodynamics and Sedimentary Evolution of Estuaries*, Springer, Dordrecht, 315–334.

Buynevich, I. V., FitzGerald, D. M. & Parolski, K. F. 1999. *Geophysical Investigation of the Nearshore Geologic Framework, Eastern Casco Bay - Reid State Park, Maine: Data Interpretation and Implications for Late Quaternary Coastal Evolution.* U.S. Geological Survey Open-File Report **99-0380**.

Buynevich, I. V., FitzGerald, D. M. & van Heteren, S. 2004. Sedimentary records of intense storms in Holocene barrier sequences, Maine, USA. *Marine Geology,* **210**, 135–148.

Buynevich, I. V., FitzGerald, D. M. & Goble, R. J. 2007. A 1500 yr record of North Atlantic storm activity based on optically dated relict beach scarps. *Geology*, **35**, 543–546.

Carter, R. W. G., Orford, J. D., Forbes, D. L. & Taylor, R. B. 1990. Morphosedimentary development of drumlin-flank barriers with rapidly rising sea level, Story Head, Nova Scotia. *Sedimentary Geology*, **69**, 117–138.

Carter, R. W. G., Orford, J. D., Jennings, S. C., Shaw, J. & Smith, J. P. 1992. Recent evolution of a paraglacial estuary under conditions of rapid sea-level rise: Chezzetcook Inlet, Nova Scotia. *Proceedings of the Geological Association of London*, **103**, 167–185.

Carter, R. W. G., Forbes, D. L., Jennings, S. C., Orford, J. D., Shaw, J. & Taylor, R.B. 1989. Barrier and lagoon coast evolution under differing relative sea-level regimes: examples from Ireland and Nova Scotia. *In:* Ward, L.G. & Ashley, G. M. (eds) Physical Processes and Sedimentology of Siliciclastic-Dominated Lagoonal Systems. *Marine Geology*, **88,** 221–242.

Church, J.A, Clark, P.U., Cazenave, A., Gregory, J.M., Jevrejeva, S., Levermann, A., Merrifield, M.A., Milne, G.A., Nerem, R.S., Nunn, P.D., Payne, A.J., Pfeffer, W.T., Stammer, D., & Unnikrishnan, A.S. 2014. Sea Level Change *In:* Stocker, T.F., Qin, D., Plattnerm D.-K., Tignor, M.M.B., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., & Midgley, P.M. (eds) *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, Cambridge University Press, Cambridge.

Church, M., & Ryder, J. M. 1972. Paraglacial sedimentation: a consideration of fluvial processes conditioned by glaciation. *Geological Society of America Bulletin*, **83**, 3059–3072.

Chute, N. E., & Nichols, R. L. 1941. *The Geology of the Coast of Northeastern Massachusetts,* Massachusetts Department of Public Works / United States Geological Survey Cooperative Geology Project, Bulletin **7**.

COSTAS, S., & FITZGERALD, D. 2011. Sedimentary architecture of a spit-end (Salisbury Beach, Massachusetts): The imprints of sea-level rise and inlet dynamics. *Marine Geology*, **284**, 203–216.

CRIDER, H. B. 1998. *Late Pleistocene development of the Brunswick, Maine sand plain and adjacent paleochannel.* PhD thesis, Boston University.

DALY, R. A. 1934. *The Changing World of the Ice Age*. Yale University Press, New Haven.

DENTON, G. H. 2011. East Antarctic retreat. *Nature Geoscience*, **4**, 135–136.

DICKSON, S. M. 1999. *The role of storm-generated combined flows in shoreface and inner continental shelf sediment erosion, transport, and deposition.* PhD thesis, School of Marine Sciences, University of Maine.

DOLAN, R. & DAVIS, R. E. 1992. An intensity scale for Atlantic coast northeast storms. *Journal of Coastal Research*, **8**, 840–853.

DOUGHERTY, A. J., FITZGERALD, D. M., PENDLETON, E. & BUYNEVICH, I. V. 2004. Evidence for Storm-Dominated early Progradation of Castle Neck Barrier, Massachusetts, USA. *Marine Geology*, **210**, 123–134

DOUGHERTY, A. J., FITZGERALD, D. M., JOHNSTON, S. J. & FERULLO, B. 2003. Ground-Penetrating Radar Provides a Detailed View of the Complexities of Barrier Progradation, Castle Neck, Massachusetts, USA. *ASCE Special Publication*, **5**.

DOWDESWELL, J. A., OTTESEN, D. & RISE, L. 2010. Rates of sediment delivery from the Fennoscandian Ice Sheet through an ice age. *Geology*, **38**, 3–6.

Duffy, W., Belknap, D. F. & Kelley, J. T. 1989. Morphology and stratigraphy of small barrier-lagoon systems in Maine. *Marine Geology*, **88**, 243–262.

Ehlers, J. & Gibbard, P. 2008. Extent and chronology of Quaternary glaciation. *Episodes*, **31**, 211–218.

Farrell, S. C. 1969. Growth cycle of a small recurved spit, Plum Island, Massachusetts. *In:* University of Massachusetts Coastal Research Group (ed.) *Coastal Environments: Northeastern Massachusetts and New Hampshire.* University of Massachusetts, Geology Department , Amherst, Field Trip Guidebook 1969 (Contribution CG-1), 316–336.

Farrell, S. C. 1970. *Sediment Distribution and Hydrodynamics, Saco River and Scarborough Estuaries, Maine*. Coastal Research Center, University of Massachusetts, Amherst, Contribution **6**.

Farrell, S. C. 1972. Present coastal processes, recorded changes, and the post-Pleistocene geologic record of Saco Bay, Maine. PhD thesis, University of Massachusetts, Amherst.

Fenster, M. S. & FitzGerald, D. M. 1996. Morphodynamics, stratigraphy, and sediment transport patterns of the Kennebec River estuary, Maine, USA. *Sedimentary Geology*, **107**, 99–120.

Fenster, M. S., FitzGerald, D. M., Kelley, J. T., Belknap, D. F., Buynevich, I. V. & Dickson, S. M. 2001. Net ebb sediment transport in a rock-bound, mesotidal estuary during spring freshet conditions: Kennebec River estuary, Maine. *Geological Society of America Bulletin*, **113**, 1522–1531.

FitzGerald, D. M. & van Heteren, S. 1999. Classification of paraglacial barrier systems: coastal New England, USA. *Sedimentology*, **46**, 1083–1108.

FitzGerald, D. M., Rosen, P. S. & van Heteren, S. 1994. New England barriers. *In:* Davis, R. A. Jr. (ed.) *Geology of Holocene Barrier Island Systems*, Springer-Verlag, Berlin, 305–394.

FitzGerald, D. M., Buynevich, I. V., Fenster, M. S. & McKinlay, P.A. 2000. Sand dynamics at the mouth of a rock-bound, tide-dominated estuary. *Sedimentary Geology*, **131**, 25–9.

FitzGerald, D. M., Buynevich, I. V., Davis, R. A., Jr. & Fenster, M. S. 2002. New England tidal inlets with special reference to riverine-associated inlet systems. *Geomorphology*, **48**, 179–208.

FitzGerald, D. M., Buynevich, I. V., Fenster, M. S., Kelley, J. T. & Belknap, D. F. 2005. Coarse-grained sediment transport in northern New England Estuaries: A synthesis. *In:* FitzGerald, D. M. & Knight, J. (eds.) *High Resolution Morphodynamics and Sedimentary Evolution of Estuaries*. Springer, New York, 195–214.

FitzGerald, D. M., Fenster, M. S., Argow, B. A. & Buynevich, I. V. 2008. Coastal impacts due to sea-level rise. *Annual Reviews of Earth and Planetary Sciences*, **36**, 601–647.

Forbes, D.L. 2005. Paraglacial Coasts. *In:* Schwartz, M. L. (ed.) *Encyclopedia of Coastal Science*. Springer, Dordrecht, 760–762.

Forbes, D.L. & Syvitski, J.P.M. 1994. Paraglacial coasts *In:* Carter, R. W. G. & Woodroffe, C. D. (eds.) *Coastal Evolution: Late Quaternary Shoreline Morphodynamics.* Cambridge University Press, Cambridge, 373–424.

Forbes, D. L. & Taylor, R. B. 1987. Coarse-grained beach sedimentation under paraglacial conditions, Canadian Atlantic coast. *In:* FitzGerald, D. M. & Rosen, P. S. (eds) *Glaciated Coasts*. Academic Press, New York, 51–86.

Forbes, D. L., Taylor, R. B., Orford, J. D., Carter, R. W. G. & Shaw, J. 1991. Gravel barrier migration and overstepping. *Marine Geology*, **97**, 305–313.

FORBES, D. L., ORFORD, J. D., CARTER, R. W. G., SHAW, J. & JENNINGS, S. C. 1995. Morphodynamic evolution, self-organization, and instability of coarse-clastic barriers on paraglacial coasts. *Marine Geology*, **126**, 63–85.

GOLDSMITH, R. 1991. Stratigraphy of the Milford-Dedham Zone, eastern Massachusetts: An Avalonian terrane. *In:* HATCH, N.L., JR. (ed.) *The Bedrock Geology of Massachusetts*: U.S. Geological Survey Professional Paper **1366–E–J.**

GOODBRED, S.L. 2003. Response of the Ganges dispersal system to climate change: a source-to-sink view since the last interstade. *Sedimentary Geology*, **121**, 239–258.

HANSON, L. S. & CALDWELL, D. W. 1989. The Lithologic and Structural Controls on the Geomorphology of the Mountainous Areas in North-Central Maine. *In:* TUCKER, R. D. & MARVINNEY, R. G. (eds) *Studies in Maine Geology - Vol. 5: Quaternary Geology*. Maine Geological Survey, Augusta, 147–167.

HAYES, M. O. & RUBY, C. H. 1994. Barriers of Pacific Alaska. *In:* DAVIS, R. A. JR. (ed.) *Geology of Holocene Barrier Island Systems*, Springer-Verlag, Berlin, 395–433.

HEIN, C. J. 2012. *Styles of coastal evolution in response to Holocene changes in sea level and sediment supply.* PhD thesis, Boston University.

HEIN, C. J., FITZGERALD, D. & BARNHARDT, W. 2007. Holocene reworking of a sand sheet in the Merrimack Embayment, Western Gulf of Maine. *Journal of Coastal Research*, SI 50, 863–867.

HEIN, C. J., FITZGERALD, D. M., STONE, B. D., CARRUTHERS, E. A. & GONTZ, A. M. 2011a. The role of backbarrier infilling in the formation of barrier island systems. *In:* KRAUS, N. C. & ROSATI, J. D. (eds.) *Coastal Sediments '11, Proceedings of the 8th International Symposium on Coastal Engineering and Science of Coastal Sediment Processes*, 1203–1216.

HEIN, C. J., FITZGERALD, D. M., MILNE, G. A., BARD, K. A. & FATTOVICH, R., 2011b. Evolution of a Pharaonic harbor on the Red Sea: Implications for coastal response to changes in sea level and climate, *Geology*, **39**, 687–690.

HEIN, C. J., FITZGERALD, D. M., CARRUTHERS, E. A., STONE, B. D., BARNHARDT, W. A. & GONTZ, A.M. 2012. Refining the model of barrier island formation along a paraglacial coast in the Gulf of Maine. *Marine Geology*, **307**–**310**, 40–57.

HEIN, C. J., FITZGERALD, D. M., BARNHARDT, W. A. & STONE, B. D. 2013. *Onshore-Offshore Surficial Geologic Map of the Newburyport East and Northern Half of the Ipswich Quadrangles, Massachusetts*. Massachusetts Geological Survey, Geologic Map **13–01**. Scale 1:24,000.

HINE, A. C. & BOOTHROYD, J. C. 1978. Morphology, processes, and recent sedimentary history of a glacial-outwash plain shoreline, southern Iceland. *Journal of Sedimentary Research*, **48**, 901–920.

HERON, S. D., JR., MOSLOW, T. F., BERELSON, W. M., HERBERT, J. R., STEELE, G. A., III & SUSMAN, K. R. 1984. Holocene sedimentation of a wave-dominated barrier-island shoreline: Cape Lookout, North Carolina. *In*: GREENWOOD, B. & DAVIS, R. A., JR. (eds.) Hydrodynamics and Sedimentation in Wave-Dominated Coastal Environments, *Marine Geology*, **60**, 413–434.

HEWITT, K., BYRNE, M. -L., ENGLISH, M. & YOUNG, G. (eds.) 2012. *Landscapes of Transition: Landform and Sediment Associations in Cold Regions*. Kluwer, Amsterdam.

HILL, H. W., KELLEY, J. T., BELKNAP, D. F. & DICKSON, S. M. 2004. The effects of storms and storm-generated currents on sand beaches in Southern Maine, USA. *Marine Geology*, **210**, 149–168.

HUBBARD, D. K. 1976. Changes in Inlet Offset Due to Stabilization. *In:* ASCE (ed.) *Proceedings of the 15th International Conference on Coastal Engineering*, ASCE, New York, 1805–1823.

JACOBSON, G. L. & JACOBSON, H. A. 1989. An inventory of distribution and variation in salt marshes from different settings along the Maine coast. *In:* ANDERSON, W. A. & BORNS, H.W. (eds) *Neotectonics of Maine*. Maine Geological Survey Bulletin, **40**, 69–83.

JENSEN, R.E. 1983. *Atlantic Coast Hindcast, Shallow-Water, Significant Wave Information*. USACE, Vicksburg, WIS Report **9**.

KELLEY, A. R., KELLEY, J. T., BELKNAP, D. F. & GONTZ, A. M. 2011. Coastal and terrestrial impact of the isostatically forced late Quaternary drainage divide shift, Penobscot and Kennebec Rivers, Maine, USA. *Journal of Coastal Research*, **27**, 1085–1093.

KELLEY, J. T., 1987. Sedimentary environments along Maine's estuarine coastline. *In:* FITZGERALD, D. M. & ROSEN, P.S. (eds) *Glaciated Coasts*. Academic Press, New York, 151–176.

KELLEY, J. T. 1989. A Preliminary Mineralogical Analysis of Glaciomarine Mud from the Western Margin of the Gulf of Maine. *Northeastern Geology*, **11**, 141–150.

KELLEY, J. T., SHIPP, R. C. & BELKNAP, D. F. 1989. Geomorphology and Late Quaternary evolution of the Saco Bay region. *In:* TUCKER, R. D. & MARVINNEY, R. G. (eds) *Studies in Maine Geology - Vol. 5: Quaternary Geology*. Maine Geological Survey, Augusta, 47–65.

KELLEY, J. T., DICKSON, S. M., BELKNAP, D. F. & STUCKENRATH, R., JR. 1992. Sea-level change and late Quaternary sediment accumulation on the southern Maine inner continental shelf. *In:* FLETCHER, C. & WEHMILLER, J. (eds) *Quaternary Coasts of the United States: Marine and Lacustrine Systems*, Society for Sedimentary Geology, Special Publication **48**, 23–34.

KELLEY, A.R., KELLEY, J. T., BELKNAP, D. F. & GONTZ, A.M. 2011. Coastal and terrestrial impact of the isostatically forced Late Quaternary drainage divide shoft, Penobscot and Kennebec Rivers, Maine, USA. *Journal of Coastal Research*, **27,** 1085–1093.

Kelley, J.T., Kelley, A.R. & Apollonio, S. 1995a. Landforms of the Gulf of Maine. *In:* Conkling, P. (ed.) *From Cape Cod to the Bay of Fundy: An Environmental Atlas of the Gulf of Maine*, The MIT Press, Cambridge, 19–39.

Kelley, J. T., Gehrels, W. R. & Belknap, D. F. 1995b. Late Holocene relative sea-level rise and the geological development of tidal marshes at Wells, Maine, U.S.A. *Journal of Coastal Research*, **11,** 136–153.

Kelley, J. T., Belknap, D. F., FitzGerald, D. M., Barber, D. C., Dickson, S. M., van Heteren, S., Fink, L. K., Jr. & Manthorp, P. A. 1995c. *A Sand Budget for Saco Bay, Maine: Report to Maine and New Hampshire Sea Grant Programs*. Augusta, Maine Geological Survey, Open File Report **95–1**.

Kelley, J. T., Dickson, S. M., Belknap, D. F., Barnhardt, W. A. & Barber, D. C. 2003. Distribution and volume of sand bodies on the rocky, glaciated inner continental shelf of the northwestern Gulf of Maine. *Journal of Coastal Research*, **19**, 41–56.

Kelley, J. T., Barber, D. C., Belknap, D. F., FitzGerald, D. M., van Heteren, S. & Dickson, S. M. 2005. Sand budgets at geological, historical and contemporary time scales for a developed beach system, Saco Bay, Maine, USA. *Marine Geology*, **214**, 117–142.

Kelley, J. T., Belknap, D. F. & Claesson, S. H. 2010. Drowned coastal deposits with associated archaeological remains from a sea-level "slowstand": Northwestern Gulf of Maine, USA. *Geology*, **38**, 695–698.

Kelley, J. T., Belknap, D. F., Kelley, A. R. & Claesson, S. H. 2013. A model for drowned terrestrial habitats with associated archeological remains in the northwestern Gulf of Maine, USA. *Marine Geology*, **338**, 1–16.

Kirwan, M. L., Murray, A. B., Donnelly, J. P. & Corbett, D. R. 2011. Rapid wetland expansion during European settlement and its implication for marsh survival under modern sediment delivery rates, *Geology*, **39**, 507–510.

Kliewe, H. & Janke, W. 1991, Holozäner Küstenausgleich im südlichen Ostseegebiet bei besonderer Berücksichtigung der Boddenausgleichsküste Vorpommerns, *Petermanns Geographische Mitteilungen*, **135**, 1–15.

Knight, J. & Harrison, S. 2009. Periglacial and paraglacial environments: a view from the past into the future. *In:* Knight, J. & Harrison, S. (eds) *Periglacial and Paraglacial Processes and Environments*. Geological Society, London, Special Publications, **320**, 1–4.

Leeder, M. R., Harris, T. & Kirkby, M. J. 1998. Sediment supply and climate change: implications for basin stratigraphy. *Basin Research*, **10**, 7–18.

Lentz, S. 2012. Buoyant coastal currents. *In:* Chassignet, E., Cenedese, C. & Verron, J. (eds) *Buoyancy-Driven Flows.* Cambridge University Press, Cambridge, 164–202.

Lynch, D. R., Holboke, M. J. & Nairrie, C. E. 1997. The Maine coastal current: spring climatological circulation. *Continental Shelf Research*, **17**, 605–634.

Lyons, J.B., Bothner, W.A., Moench, R.H. & Thompson, J.B. 1997. *Bedrock Geology Map of New Hampshire*, U.S. Geological Survey, scale 1:500,000.

McCormick, C.L. 1968. Holocene stratigraphy of the marshes at Plum Island, Massachusetts, PhD thesis, University of Massachusetts.

McIntire, W. G. & Morgan, J. P. 1964. *Recent Geomorphic History of Plum Island, Massachusetts, and Adjacent Coasts*. Louisiana State University Press, Baton Rouge, Louisiana State University Coastal Studies Series, **8**.

MERCIER, D. 2009. Paraglacial and paraperiglacial landsystems: concepts, temporal scales and spatial distribution. *Géomorphologie*, **4/2008**, 223–233.

MILLIMAN, J. D. & FARNSWORTH, K. L. 2011. *River Discharge to the Coastal Ocean.* Cambridge University Press, Cambridge, 384 p.

NICHOL, S. L. & BOYD, R. 1993. Morphostratigraphy and facies architecture of sandy barriers along the Eastern Shore of Nova Scotia. *Marine Geology*, **114,** 59–80.

NUMMEDAL, D., HINE, A. C. & BOOTHROYD, J. C. 1987. Holocene evolution of the south-central coast of Iceland. *In:* FITZGERALD, D. M. & ROSEN, P. R. (eds.) *Glaciated Coasts*. Academic Press, New York, 115–150.

OLDALE, R. N. 1985. A drowned Holocene barrier spit off Cape Ann, Massachusetts. *Geology*, **13**, 375–377.

OLDALE, R. N. 1989. Timing and mechanisms for the deposition of the glacimarine mud in and around the Gulf of Maine: a discussion of alternative models. *In:* TUCKER, R. D. & MARVINNEY, R. G. (eds) *Studies in Maine Geology - Vol. 5: Quaternary Geology*. Maine Geological Survey, Augusta, 1–10.

OLDALE, R. N. & EDWARDS, G. B. 1990. *Cores From Marine Geologic Features in the Western Gulf of Maine*. U.S. Geological Survey Field Study Map, **MF-2147**.

OLDALE, R. N., WOMMACK, L. E. & WHITNEY, A. B. 1983. Evidence for a postglacial low relative sea-level stand in the drowned delta of the Merrimack River, western Gulf of Maine, *Quaternary Research*, **19**, 325–336.

OLDALE, R. N., COLMAN, S. M. & JONES, G. A. 1993. Radiocarbon ages from two submerged strandline features in the western Gulf of Maine and a sea-level curve for the northeastern Massachusetts coastal region. *Quaternary Research*, **40**, 38–45.

ORFORD, J. D., & CARTER, R. W. G. 1985. Storm-generated dune armouring on a sand-gravel barrier system, southeastern Ireland. *Sedimentary Geology*, **42**, 65–82.

ORFORD, J. D., CARTER, R. W. G. & JENNINGS, S. C. 1991. Coarse clastic barrier environments: evolution and implications for Quaternary sea-level interpretation. *Quaternary International*, **9**, 87–104.

OSBERG, P.H., HUSSEY, A.M., II & BOONE, G.M. 1985. *Bedrock Geologic Map of Maine,* Maine Geological Survey, scale 1:500,000.

RAMPINO, M.R. & SANDERS, J.E. 1981. Evolution of the barrier islands of southern Long Island, New York. *Sedimentology*, **28**, 37–47.

RETELLE, M. J. & WEDDLE, T.K. 2001. Deglaciation and relative sea-level chronology, Casco Bay Lowland and lower Androscoggin River valley, Maine. *In:* WEDDLE, T. K. & RETELLE, M. J. (eds) Deglacial History and Relative Sea-Level Changes, Northern New England and Adjacent Canada, Geological Society of America Special Papers, **351**, 191–214.

RHODES, E. G. 1973. *Pleistocene-Holocene Sediments Interpreted by Seismic Refraction and Wash-Bore Sampling, Plum Island-Castle Neck, Massachusetts*, Coastal Engineering Research Center, U.S. Army Corps of Engineers, Vicksburg, USACE Technical Memorandum **40**.

RIDGE, J. C. 2004. The Quaternary glaciations of western New England with correlations to surrounding areas. *In:* EHLERS, J. & GIBBARD, P. L. (eds) *Quaternary Glaciations, Extent and Chronology Part II: North America*. Elsevier, Amersterdam, 169–199.

ROBINSON, G.R. & KAPO, K.E. 2003. *Generalized Lithology and Lithogeochemical Character of Near-Surface Bedrock in the New England Region*, U.S. Geological Survey Open File Report **03–225**.

Rosen, P. S. 1984. Gravel spit processes, Thompson Island, Boston Harbor, Massachusetts. *In:* Hanson, L. (ed.) *Geology of the Coastal Lowlands: Boston, MA to Kennebunk, ME.* Salem State College, Salem, New England Intercollegiate Geological Conference Guidebook **76**, 25–38.

Ryder, J. M. 1971. Some aspects of the morphometry of paraglacial alluvial fans in south-central British Columbia. *Canadian Journal of Earth Sciences*, **8**, 1252–1264.

Sallenger, A. H. Jr., Doran, K. S. & Howd, P. A. 2012. Hotspot of accelerated sea-level rise on the Atlantic coast of North America. *Nature Climate Change*, **2**, 884–888.

Short, A.D. 1979. Barrier island development along the Alaskan-Yukon coastal plains. *Geological Society of America Bulletin, Part II*, **90**, 77–103.

Slaymaker, O. 2009. Proglacial, periglacial or paraglacial?. *In:* Knight, J. & Harrison, S. (eds) *Periglacial and Paraglacial Processes and Environments*. Geological Society, London, Special Publications, **320**, 71–84.

Smith, J.B. 1991. *Morphodynamics and Stratigraphy of Essex River Ebb-Tidal Delta: Massachusetts.* U.S. Army Engineer Waterways Experiment Station, U.S. Army Corps of Engineers, Vicksburg, USACE, Technical Report **CERC-91-11**.

Smith, J. B., & FitzGerald, D. M. 1994. Sediment transport patterns at the Essex River Inlet ebb tidal delta, Massachusetts, U.S.A. *Journal of Coastal Research*, **10**, 752–774.

Stea, R. R., Boyd, R., Fader, G. B. J., Courtney, R. C., Scott, D. B., and Pecore, S. S. 1994. Morphology and seismic stratigraphy of the inner continental shelf off Nova Scotia, Canada: Evidence for a -65 m low-stand between 11,650 and 11,250 ^{14}C yr B.P. *Marine Geology*, **117**, 135–154.

STONE, B. D. & PEPER, J. D. 1982. Topographic control of the deglaciation of eastern Massachusetts. *In:* LARSON, G. J. & STONE, B. D. (eds) *Late Wisconsin Glaciation of New England*. Kendall/Hunt, Dubuque, 145–166.

STONE, B. D., STONE, J. R. & MCWEENEY, L. J. 2004. Where the glacier met the sea: Late Quaternary geology of the northeast coast of Massachusetts from Cape Ann to Salisbury. *In:* Hanson, L. (ed.) *Geology of the Coastal Lowlands: Boston, MA to Kennebunk, ME.* Salem State College, Salem, New England Intercollegiate Geological Conference Guidebook **76**, Trip B-3.

STONE, B. D., STONE, J. R. & DIGIACOMO-COHEN, M. L. 2006. *Surficial Geologic Map of the Salem-Newburyport East-Wilmington-Rockport Quadrangle Area in Northeast Massachusetts*, U.S. Geological Survey, Reston, Virginia, U.S. Geological Survey Open-File Report, **2006-1260-B**.

SWIFT, D. J. P. 1968. Coastal erosion and transgressive stratigraphy. *Journal of Geology*, **76**, 444–456.

SYVITSKI, J. P. 1991. Towards an understanding of sediment deposition on glaciated continental shelves. *Continental Shelf Research*, **11**, 897–937.

TARY, A. K., FITZGERALD, D. M. & BUYNEVICH, I.V. 2001. Late Quaternary morphogenesis of a marine-limit delta plain in southwest Maine. *In:* WEDDLE, T. K. & RETELLE, M. J. (eds) Deglacial History and Relative Sea-Level Changes, Northern New England and Adjacent Canada, Geological Society of America Special Papers, **351**, 125–149.

THOMPSON, W. B. & BORNS, H. W., JR. 1985. *Surficial Geologic Map of Maine*. Maine Geological Survey, Augusta. Scale: 1:500,000.

THOMPSON, W. B., CROSSEN, K. J., BORNS, H. W., JR. & ANDERSEN, B. G. 1989. Glaciomarine deltas of Maine and their relation to late Pleistocene–Holocene crustal movements. *In:* ANDERSEN, W. A. & BORNS, H. W., JR. (eds) *Neotectonics of Maine: Studies in Seismicity,* *Crustal Warping, and Sea-Level Change*: Maine Geological Survey, Augusta, Bulletin **40**, 43–67.

VAN HETEREN, S. 1996. Preserved records of coastal-morphologic and sea-level changes in the stratigraphy of paraglacial barriers. PhD thesis, Boston University.

YANG, S. L., MILLIMAN, J. D., LI, P. & XU, K. 2010. 50,000 dams later: Erosion of the Yangtze River and its delta. *Global and Planetary Change*, **75**, 14–20.

Figure and Table Captions

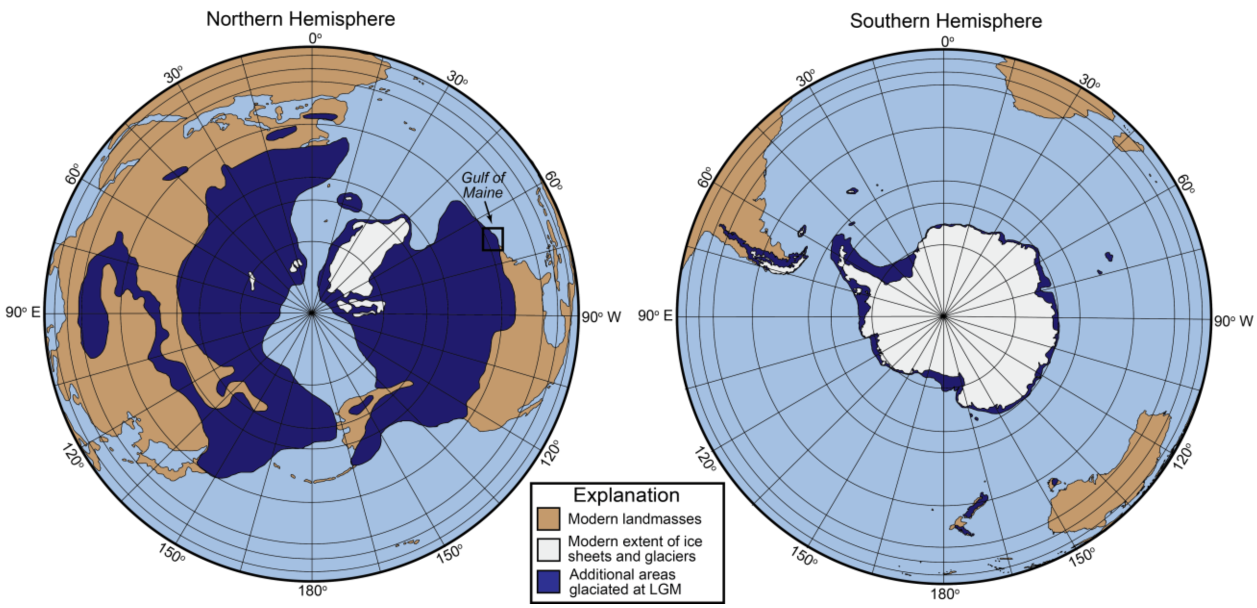


Fig. 1. Extent of formerly glaciated coasts. Northern Hemisphere map is modified from Mercier (2009); Southern Hemisphere map is constructed from data of Ehlers & Gibbard (2008) and Denton (2011). LGM: Last Glacial Maximum.

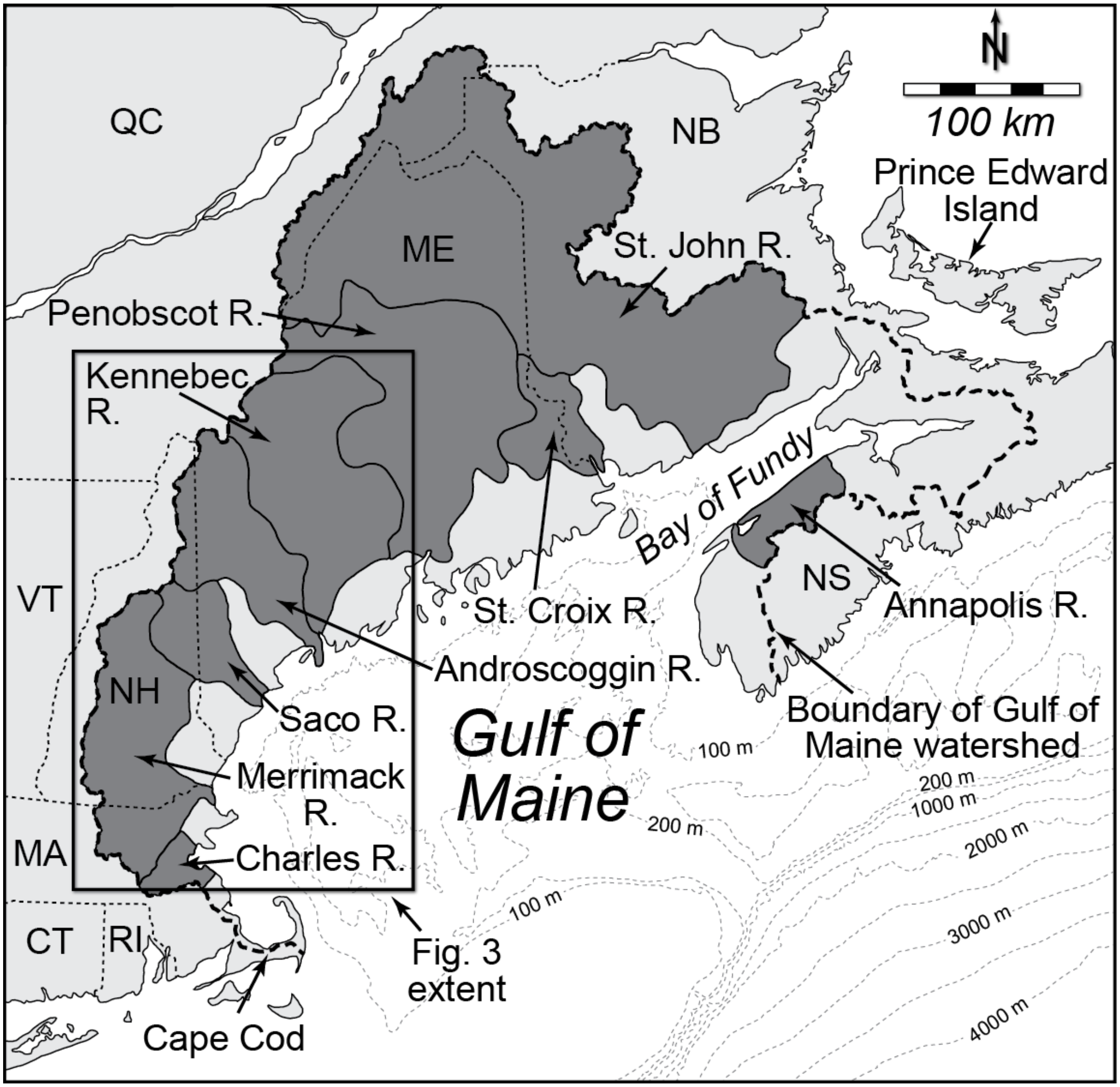


Fig. 2. Gulf of Maine and adjacent New England. Major drainage basins of the Gulf of Maine are shown in dark grey (modified from Kelley *et al.* 1995a).

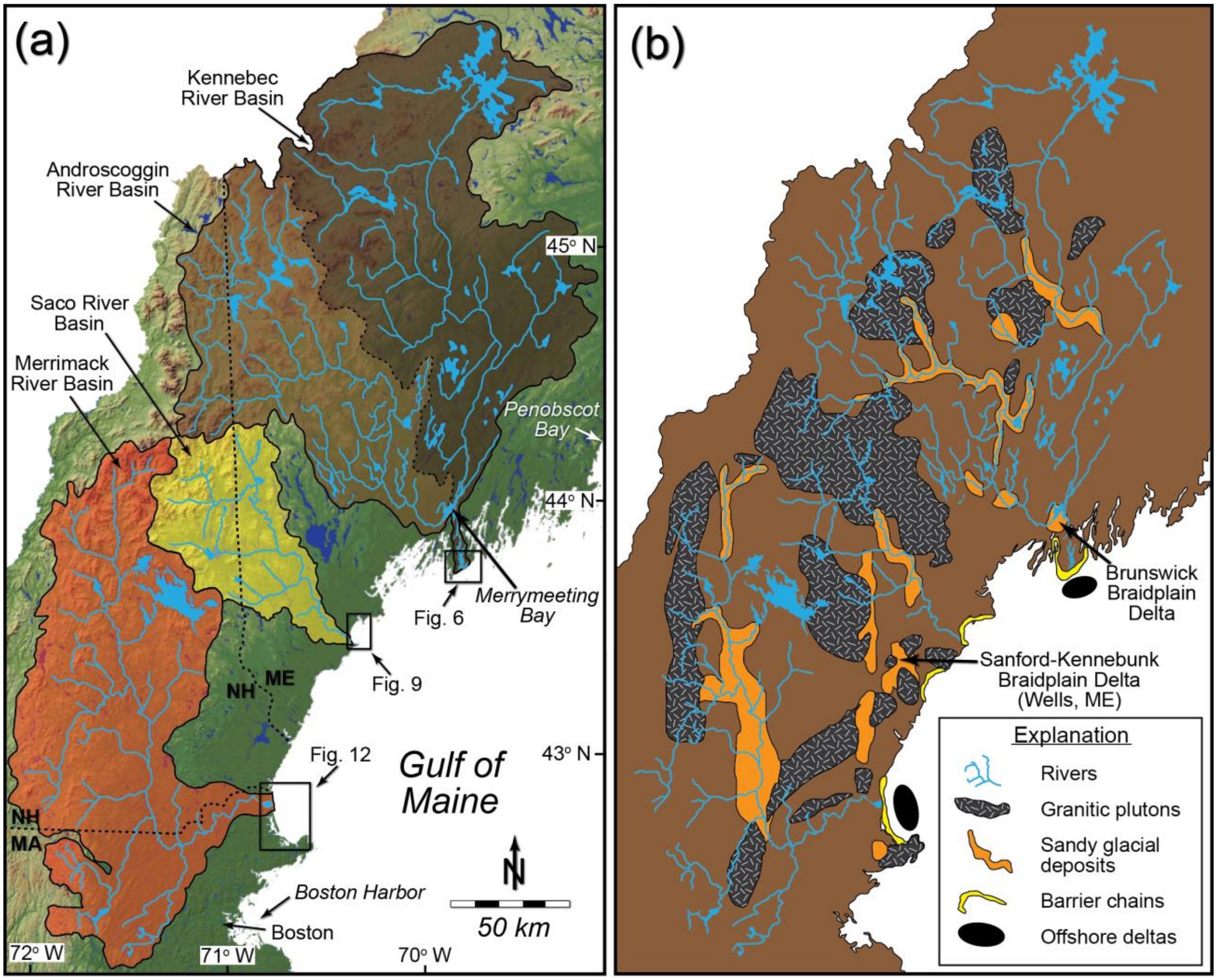


Fig. 3. (**a**) Drainage basins affiliated with the river-associated Kennebec/Androscoggin, Saco and Merrimack paraglacial barrier systems. (**b**) The distribution of plutons and sandy glacial deposits in the drainage basins and of the barriers and offshore deltas associated with these rivers (modified from FitzGerald *et al.* 2005).

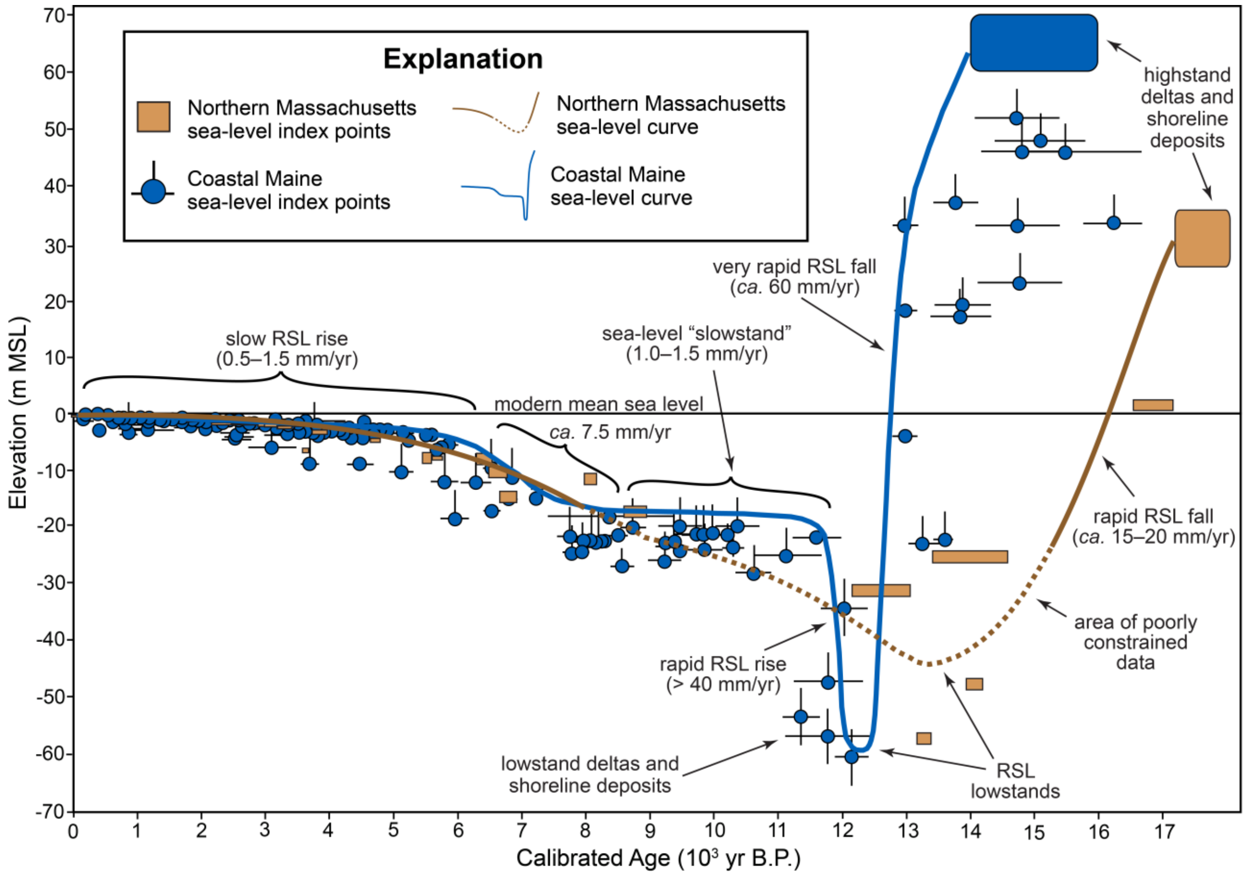


Fig. 4. Sea-level history of the western Gulf of Maine. Northern Massachusetts RSL curve is modified from Hein *et al.* (2012). Coastal Maine RSL curve is modified from Kelley *et al.* (2010).

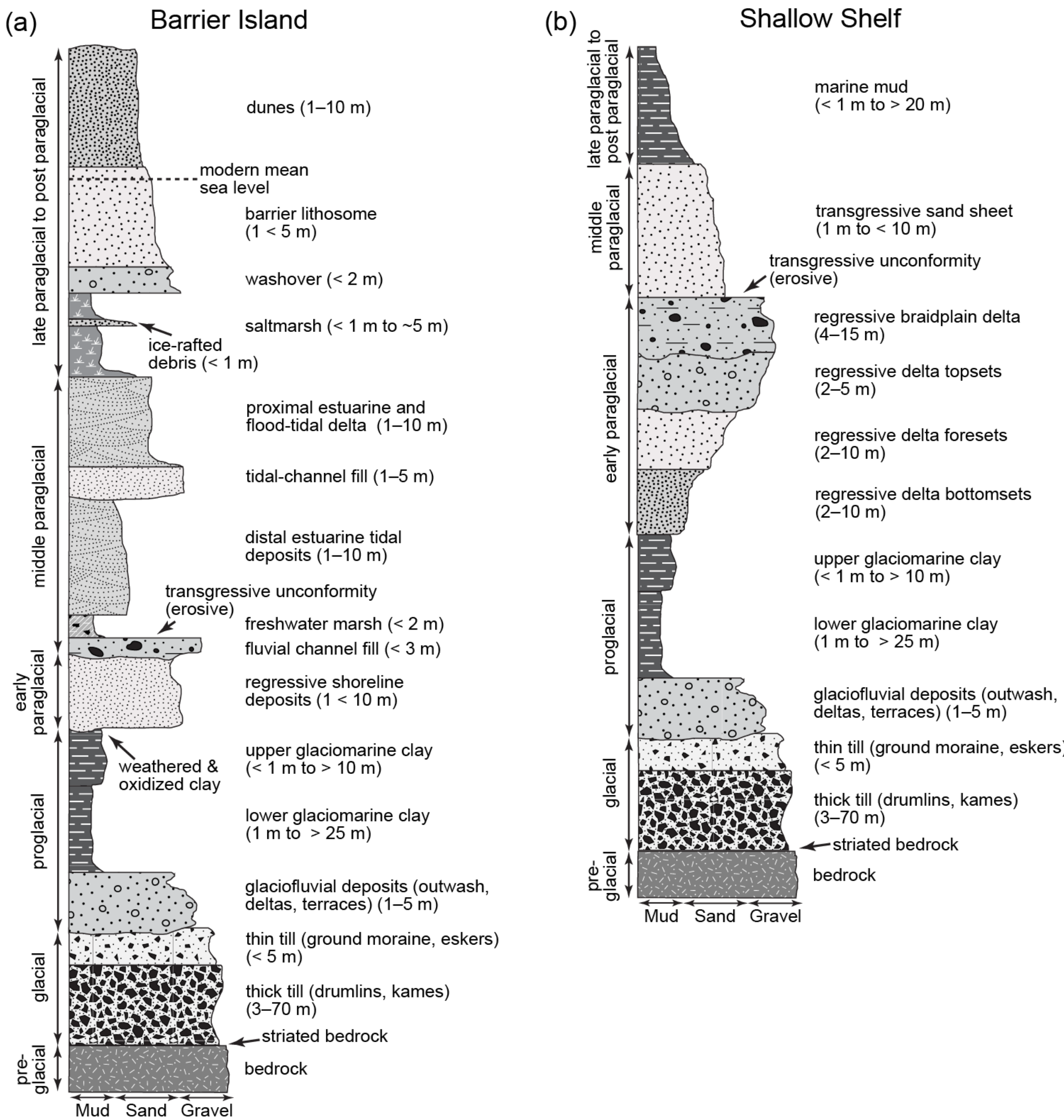


Fig. 5. Idealized stratigraphic sections through (**a**) river-associated paraglacial barrier island and (**b**) shallow-shelf palaeodelta sequence offshore a river-associated paraglacial barrier. Thicknesses given for each unit are approximate and estimated from data in McIntire & Morgan (1964), Rhodes (1973), van Heteren (1996), Buynevich (2001), Buynevich & FitzGerald (2001), Stone *et al.* (2006), Barnhardt *et al.* (2009) and Hein *et al*. (2013).

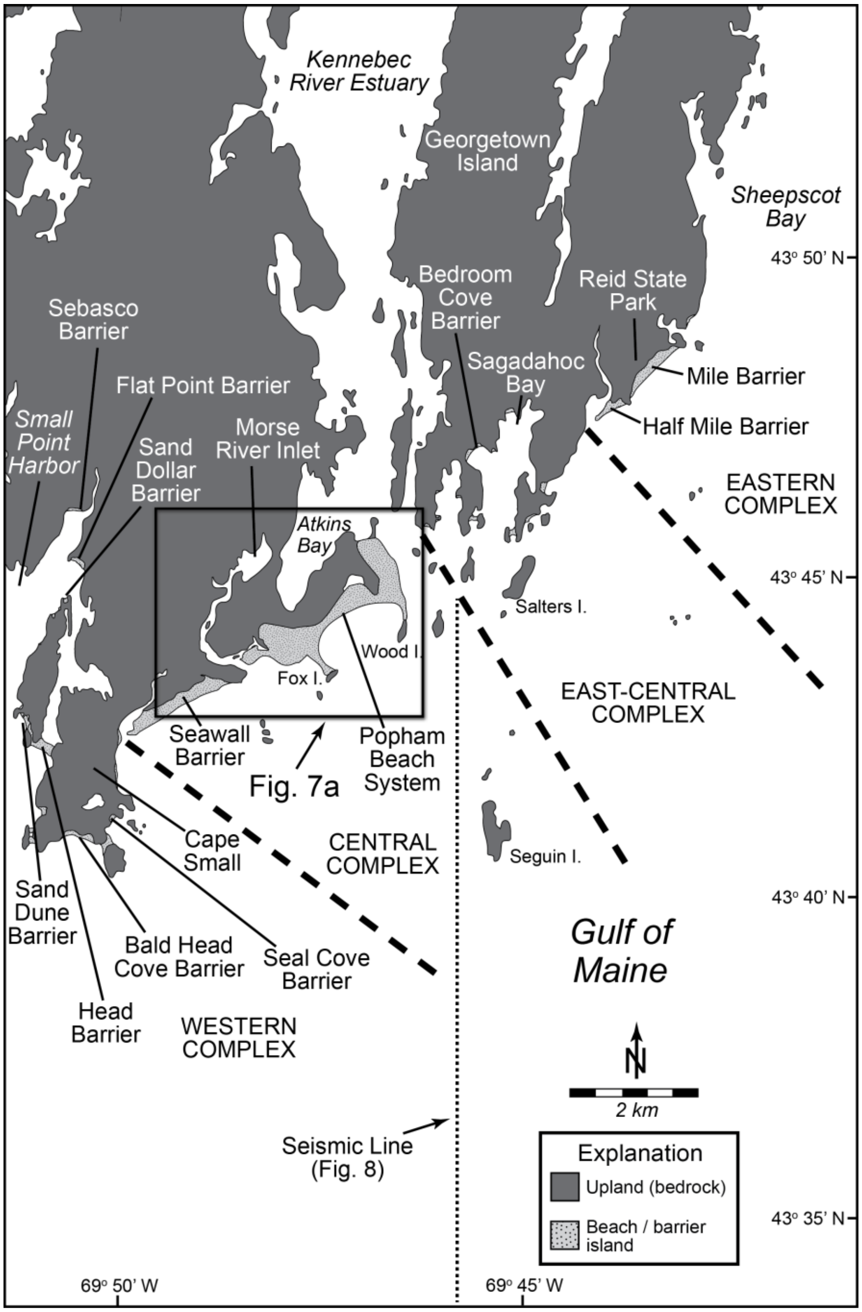


Fig. 6. Kennebec barrier chain showing sandy Holocene coastal landforms with a range of orientations. Four physiographic compartments are distinguished on the basis of morphology and sedimentology (Buynevich 2001): (1) western (Small Point Harbor / Cape Small barriers), (2) central (Seawall / Popham barriers), (3) east-central (Sagadahoc Bay barriers), and (4) eastern (Reid State Park barriers).

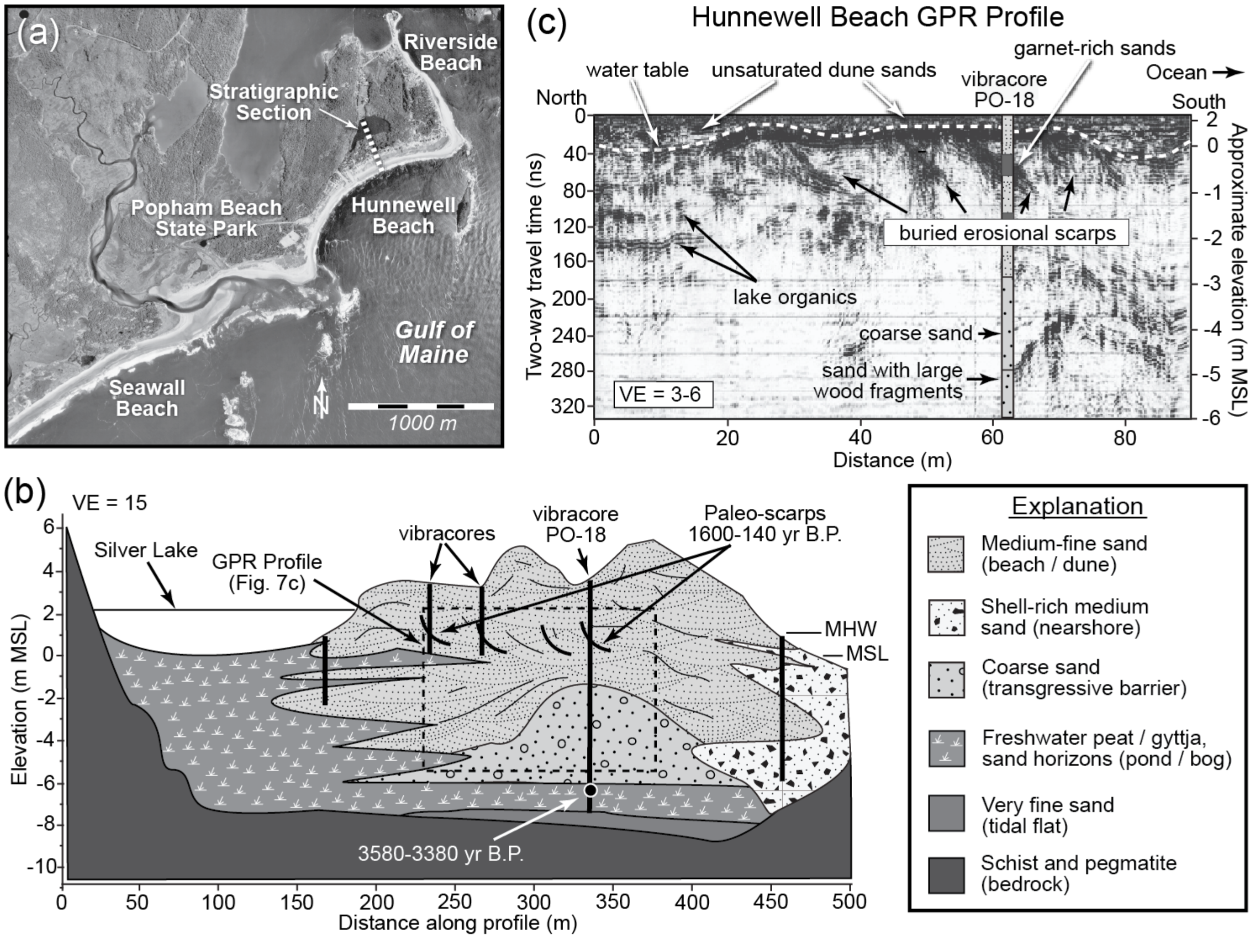


Fig. 7. Stratigraphy of Hunnewell Beach (modified from Buynevich *et al.* 2004). (**a**) Satellite image of Popham Beach complex, with Riverside, Hunnewell and Seawall beaches, showing transect location (for overall location, see Fig. 6). (**b**) Stratigraphic cross-section extending from the modern beach to Silver Lake, showing the extent of both transgressive and regressive facies, with the latter punctuated by a series of buried erosional scarps (optical chronology from Buynevich *et al.* 2007). (**c**) Analogue GSSI GPR section collected with a 200-MHz antenna and showing transgressive barrier core overlying bedrock, in turn overlain by a prograded barrier sequence. (MHW: mean high water; MSL: mean sea level; m MSL: metres with respect to mean sea level; VE: vertical exaggeration).

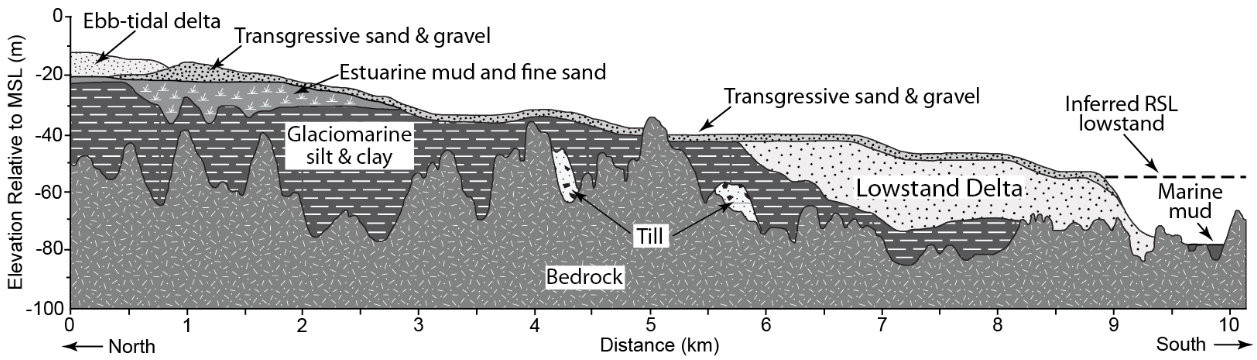


Fig. 8. Shore-normal stratigraphic cross section across Kennebec palaeodelta (modified from Barnhardt *et al.* 1997).Cross section is interpreted from high-resolution boomer seismic-reflection data, ground-truthed with four vibracores. MSL: mean sea level. Vertical exaggeration = 25x. Transect location shown in Fig. 6.

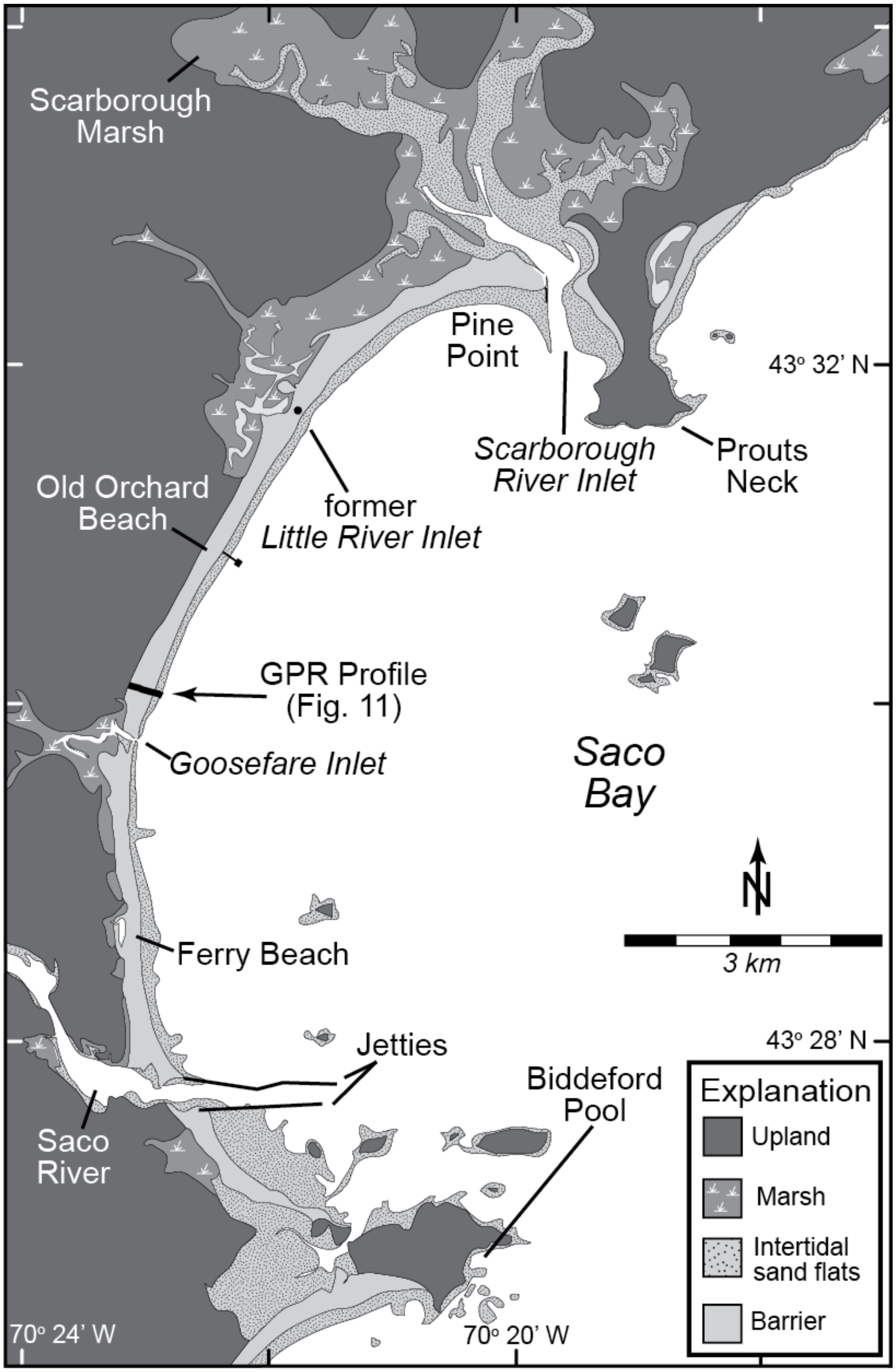


Fig. 9. Distribution of barrier and backbarrier sediments within the Saco Bay barrier chain. No lowstand delta is present in Saco Bay, owing to the trapping of glacial and paraglacial sediment in upstream river plains, lakes and wetlands.

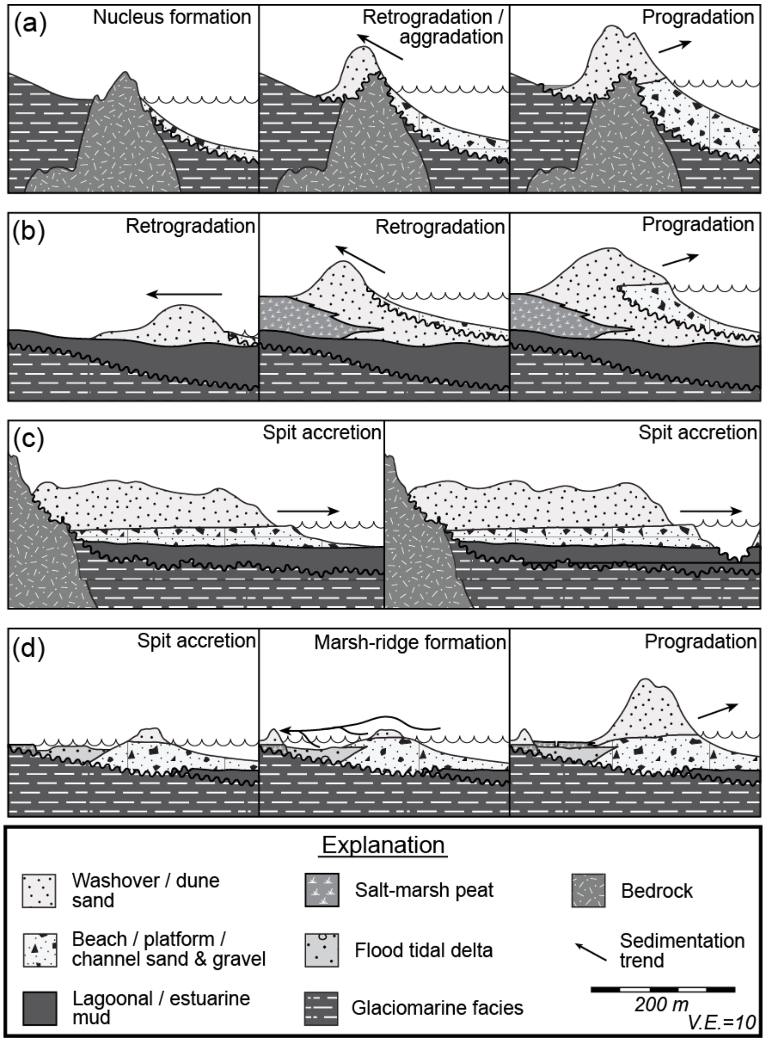


Fig. 10. Barrier-sequence types at Saco Bay, interpreted from geophysical (GPR) and sediment-core data (modified from Van Heteren 1996). (**a**) Headland-beach sequences interpreted to be the result of beach stabilisation at outlying headlands that formed pinning points for subsequent barrier-spit accretion. (**b**) Simple successions of barrier and backbarrier facies in which peat and inorganic backbarrier facies have infilled irregular palaeo-topography and are partly capped by washover and eolian sand in a retrogradational succession. (**c**) Successions of inlet-proximal barrier-spit and tidal-inlet facies in which coarse sandy and gravelly lag deposits formed in inlet channels fine upward into platform facies, covered in turn by somewhat coarser spit-beach facies and capped by aeolian sand. This sequence, with a strong shore-parallel element of variability, is

interpreted to form by longshore migration of a barrier-spit and tidal-inlet system over considerable lateral distances (*cf.* Heron *et al.* 1984). Most of the spit sequences show a northerly component of net migration, which is reflected in the shape of recurved ridges along the landward barrier margin (Kelley *et al.* 1989). (**d**) Complex juxtaposition of barrier and backbarrier facies, with marsh ridges forming near inlets during storms.

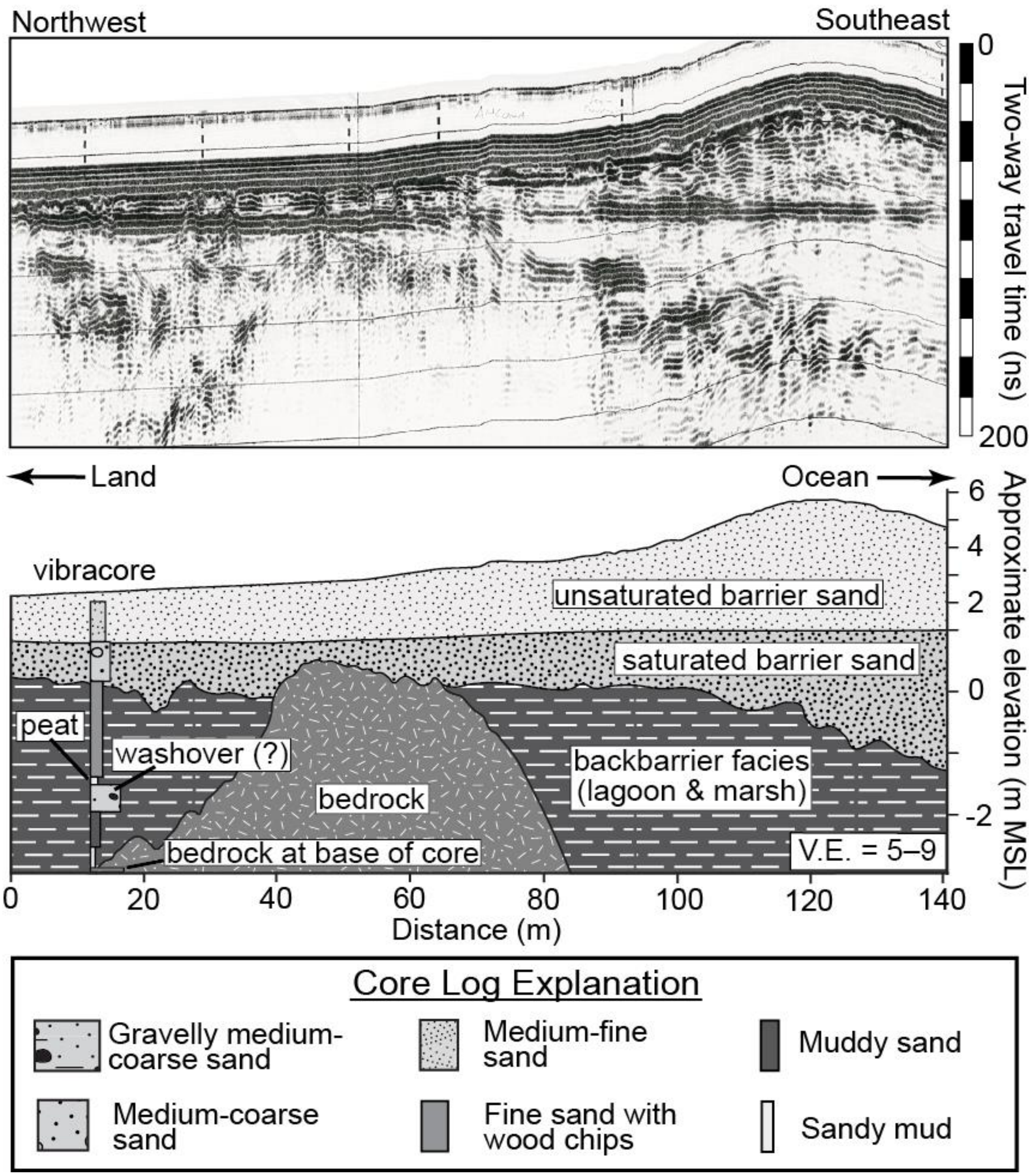


Fig. 11. Shore-perpendicular analogue GSSI GPR section across Saco barrier chain, collected with a 120-MHz antenna and showing barrier anchoring to a bedrock pinning point that is no longer recognizable at the surface. Profile location is given in Fig. 9.

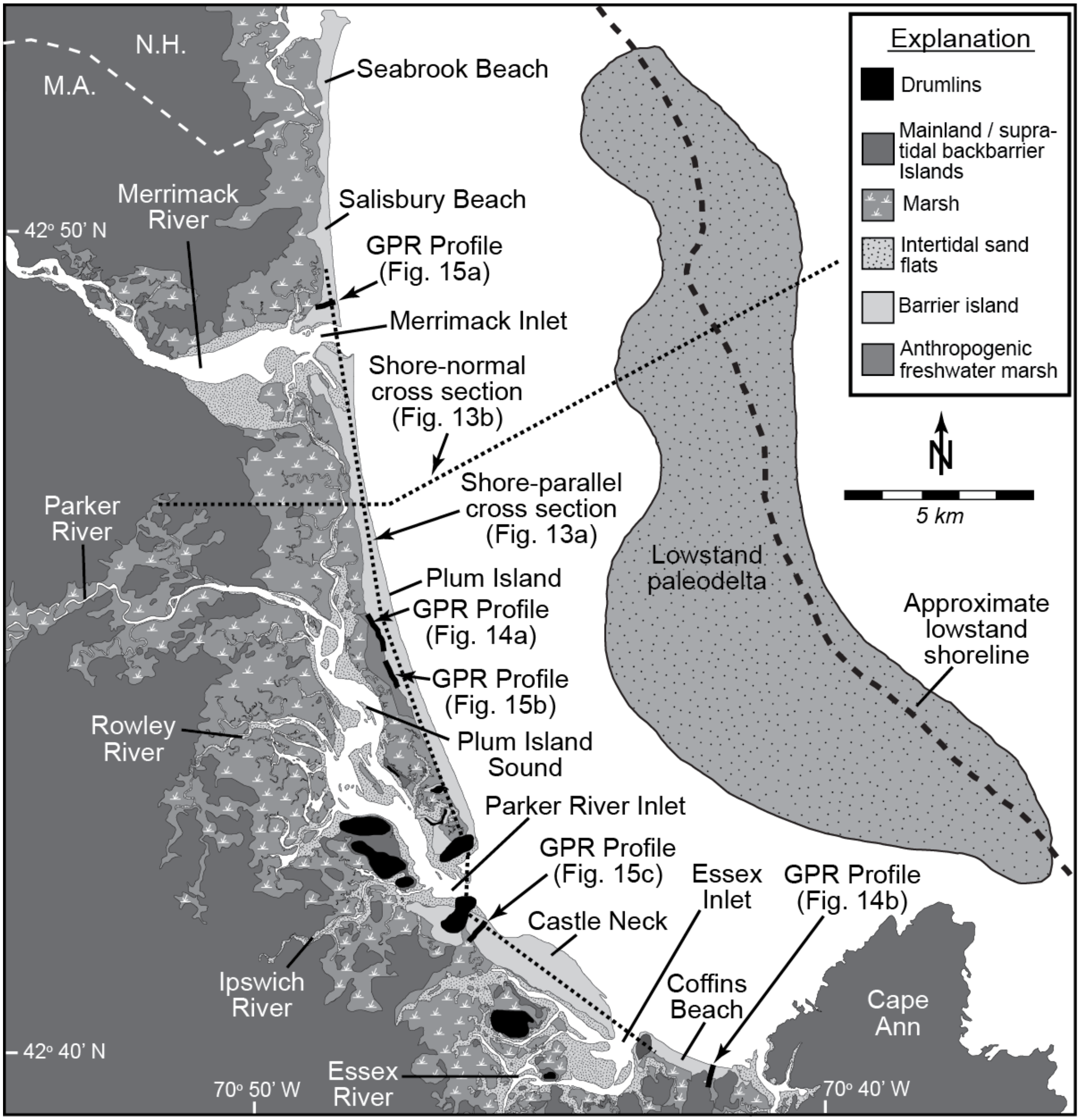


Fig. 12. Merrimack Embayment barrier chain and lowstand delta. (Modified from Hein *et al.* 2012). (Lowstand shoreline and delta locations are derived from Oldale *et al.* (1983) and Hein *et al.* (2013)).

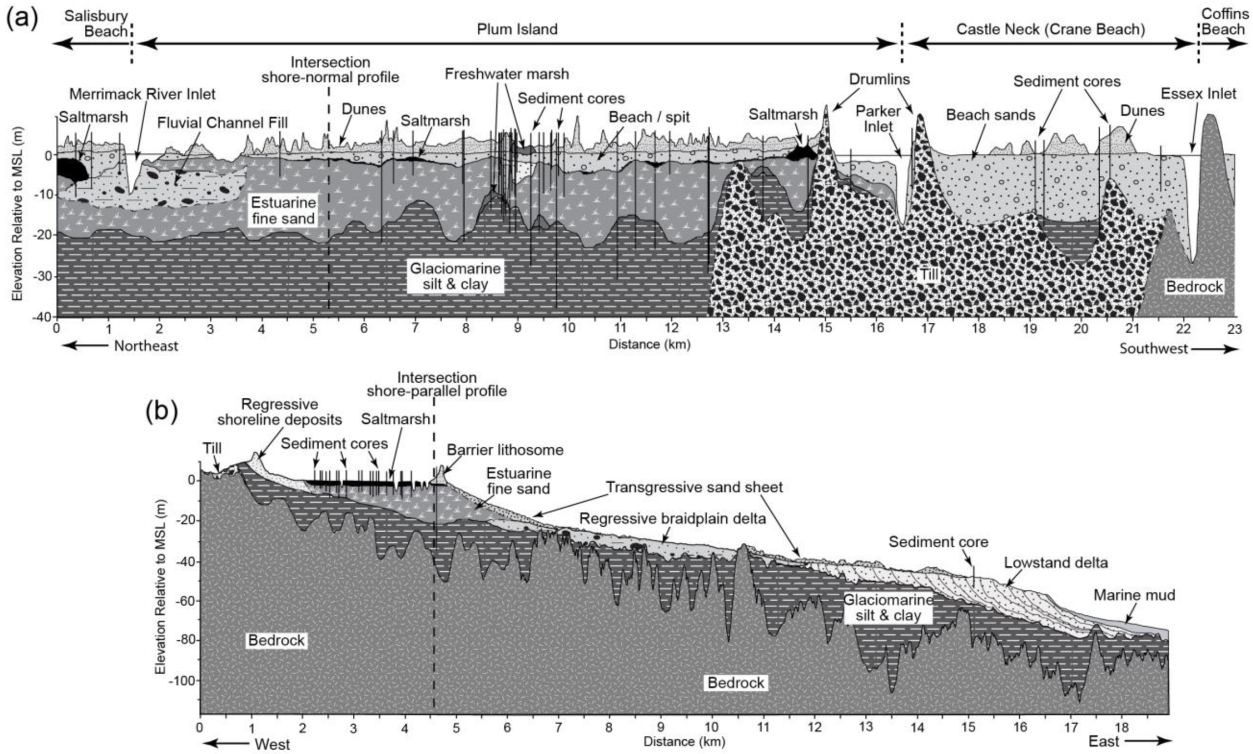


Fig. 13. Stratigraphic cross sections of barrier chain at the mouth of the Merrimack River. Locations shown in Fig. 12. (**a**) Shore-parallel cross section across four barriers in Merrimack Embayment (modified and expanded from Hein *et al.* 2013). Salisbury Beach and Plum Island sections are based on more than 20 km of GPR profiles (Hein *et al.* 2012, 2013), ground-truthed with core data from McIntire & Morgan (1964), McCormick (1968), Rhodes (1973), Costas & FitzGerald (2011), Hein (2012) and Hein *et al.* (2012). Castle Neck section of cross-section is based on cores from Rhodes (1973). (**b**) Shore-normal cross section (modified from Hein *et al.* 2013). Eastern half of cross section is based on high-resolution Chirp seismic-reflection data (Barnhardt *et al.* 2009), ground-truthed with surficial sediment samples (not shown) and one offshore vibracore. Western half is ground-truthed with core data from McIntire & Morgan (1964), McCormick (1968), Oldale & Edwards (1990) and Hein (2012). Offshore core data are courtesy of G. Edwards. MSL: mean sea level. Note that vertical exaggeration of (**a**) is exactly twice the vertical exaggeration of (**b**).

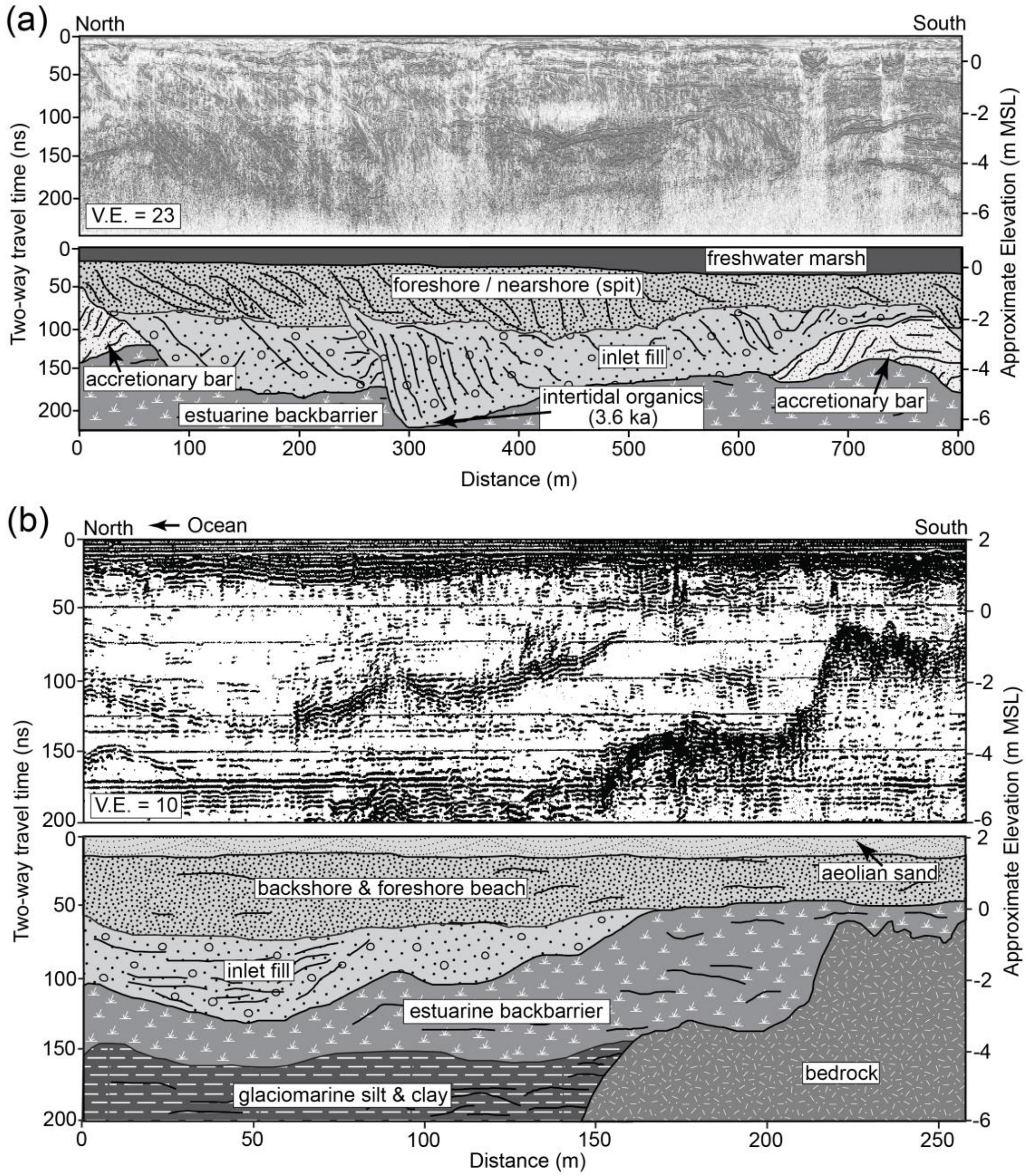
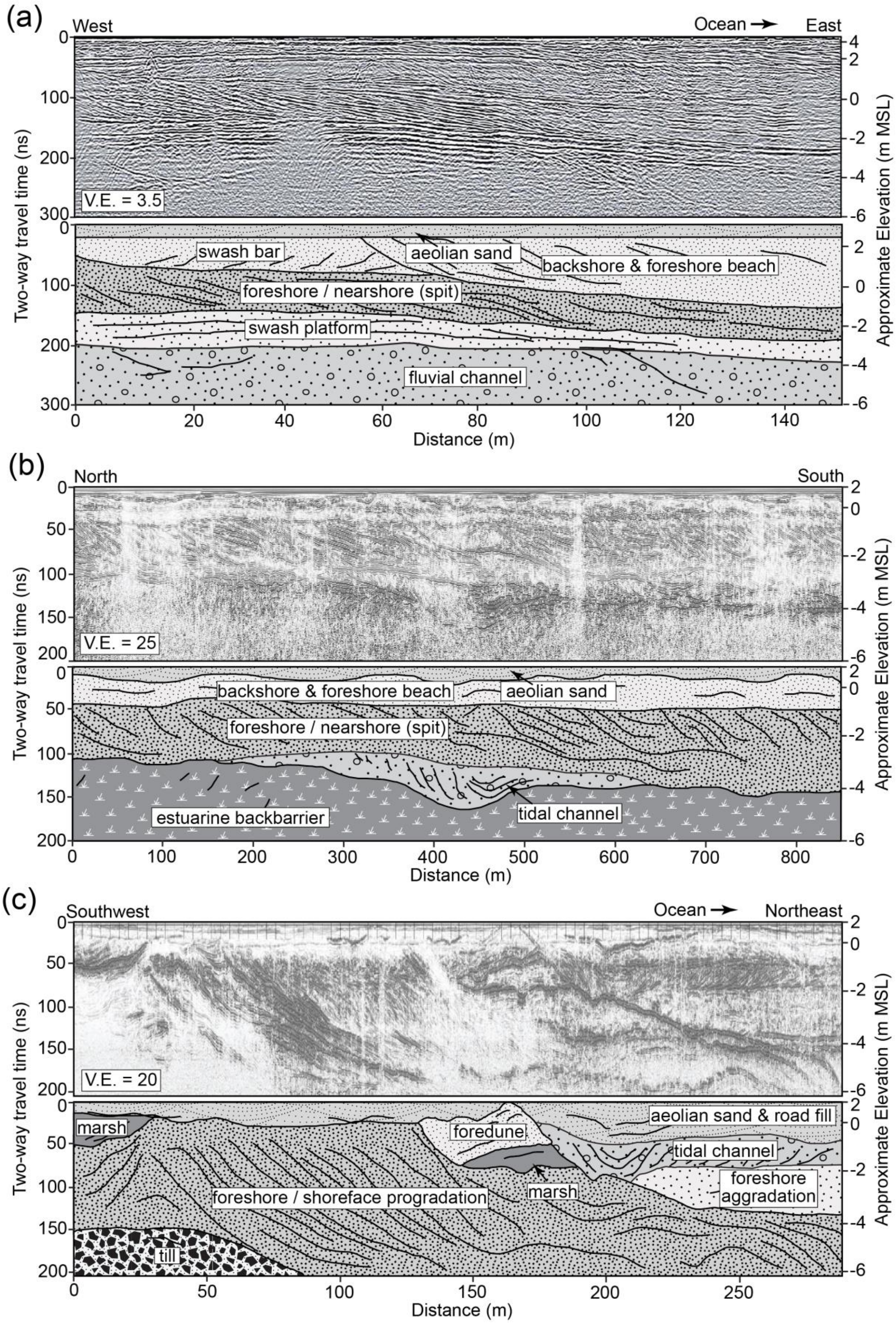


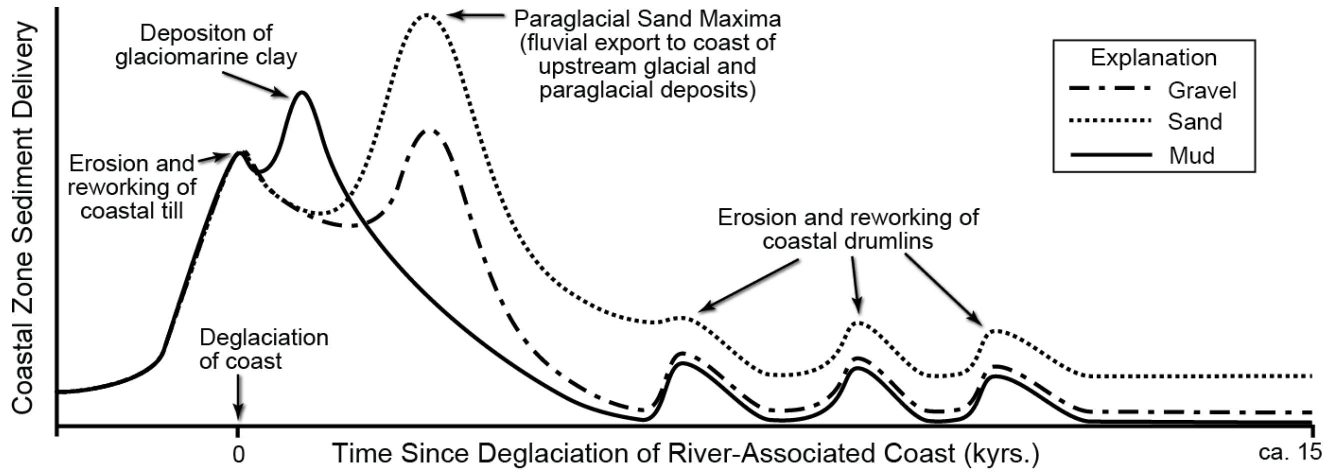
Fig. 14. Buried palaeo-inlet sequences at (**a**) Plum Island (shore-parallel) and (**b**) Coffins Beach (shore-normal), as imaged in GPR profiles. Profile locations shown in Fig. 12. Plum Island GPR profile is modified from Hein *et al.* (2013) and was collected with a digital GSSI SIR-2000 system with a 200-MHz antenna and digitally post-processed using the GSSI Radan software package. Coffins Beach GPR profile was collected with an analogue GSSI GPR with a 200-MHz antenna. (Profile is courtesy of P. McKinlay).



(a)
West
Ocean
East
V.E. = 3.5
swash bar
aeolian sand
backshore & foreshore beach
foreshore / nearshore (spit)
swash platform
fluvial channel
Two-way travel time (ns)
Approximate Elevation (m MSL)
Distance (m)
(b)
North
South
V.E. = 25
backshore & foreshore beach
aeolian sand
foreshore / nearshore (spit)
estuarine backbarrier
tidal channel
Two-way travel time (ns)
Approximate Elevation (m MSL)
Distance (m)
(c)
Southwest
Ocean
Northeast
V.E. = 20
marsh
foredune
aeolian sand & road fill
tidal channel
foreshore aggradation
marsh
foreshore / shoreface progradation
till
Two-way travel time (ns)
Approximate Elevation (m MSL)
Distance (m)

1569

Fig. 15. Representative post-processed GPR profiles demonstrating the dominant formation mechanisms of the Merrimack Embayment barriers. All GPR profiles were collected using a digital GSSI SIR-2000 system with a 200-MHz antenna. Profile locations shown in Fig. 12. (**a**) Shore-normal GPR section across Salisbury Beach, demonstrating the contribution of foreshore drift and swash-bar welding to barrier elongation. (Profile modified from Costas & FitzGerald 2011). (**b**) Shore-parallel GPR section across central Plum Island showing spit progradation over intertidal backbarrier deposits, the dominant mechanism of progradation and elongation of the barrier-spit system. (Profile modified from Hein 2012). (**c**) Shore-normal GPR section across Castle Neck containing high-amplitude reflections representative of heavy-mineral concentrations deposited during storm events (Dougherty *et al*. 2004). Here, barrier growth was dominated by seaward progradation. Seaward part of unit labelled as a “tidal channel” is interpreted as an onshore-migrating bar associated with the southward migration of the Parker River Inlet. Profile modified from Dougherty *et al.* (2003).



Period	Glacial	Proglacial	Early Paraglacial	Middle Paraglacial	Late Paraglacial	Post Paraglacial
Relative sea-level (RSL) trends	isostatically depressed crust; tide-water glaciers	isostatically depressed crust; RSL highstand	crustal rebound; rapid RSL fall & lowstand	rapid RSL rise	slow RSL rise	slow RSL rise
Dominant sediment sources	glacially transported sediments	meltwater	drainage basin glacial & paraglacial deposits	coastal paraglacial deposits (regressive shoreline and delta deposits); coastal till (e.g., drumlins, moraine)	non-glacial fluvial sediments; coastal till (drumlins, moraines)	fluvial sediments derived only from bedrock erosion
Dominant sediment-delivery mechanisms	glacier ice-contact; sub-glacial water	meltwater	meltwater; braided streams; shoreline reworking	transgressive erosion; shoreface reworking; rivers	shoreline erosion & reworking; waves, tides, rivers	shoreline erosion & reworking; waves, tides, rivers
Associated landforms / deposits	drumlins, moraines, kames, eskers	outwash, ice-front deltas, glaciomarine clay	regressive shorelines & deltas; braidplain, lowstand deltas	shallow shelf sand sheet; proto-barriers; tidal-channel fills; backbarrier estuarine deposits	stable, progradational barriers and beaches; marshes; dunes; marine mud	barriers and beaches; marshes; dunes; marine mud
Approximate timing along river-associated coasts in GoM	prior to 23.7 ka to between 18 (MA) and 16 (ME) ka	17.5–15.5 ka	15.5–12 ka	12–7 ka	7 ka to present	?
Modern examples	Antarctic ice shelves	Greenland	Southeast Iceland; Copper River Delta	Eastern Canadian coast	Gulf of Maine river-associated barriers	?

Fig. 16. Schematic diagram of the pattern of sedimentation during the paraglacial period. Note that time before present increases to the left (modern is to the right). Conceptual model builds on ideas and models proposed by Church & Ryder (1972) and Ballantyne (2002b). The post-glacial period is only possible once all glacigenic and primary paraglacial deposits have been exhausted or deeply buried and can no longer contribute to barrier development. Question marks associated with this period reflect uncertainty in the possibility that such a period is ever reached, as even indirect contributions by the cannibalization of barrier segments formed from paraglacial sediments are excluded.

Table 1. *Physiography, hydraulics, and sedimentology of western Gulf of Maine estuaries (modified and updated from FitzGerald et al. 2005).*

	Kennebec/Androscoggin	Saco	Merrimack
Estuary physiography			
Geological setting	peninsula / deep embayment	bedrock valley	drowned river valley / upper bedrock valley
Paraglacial coastal setting (FitzGerald & van Heteren 1999)	mixed-energy mainland-segmented (Type 3b)	mixed-energy mainland-segmented (Type 3b)	mixed-energy inlet-segmented (Type 4b)
Spring tidal range (m)	3.0	3.1	2.9
Shallow-water wave height (m)	0.4	0.4	0.4
Tidal prism (m^3)	101×10^6	8.1×10^6	30×10^6
Estuary type	partially to vertically mixed	partially to vertically mixed	partially to vertically mixed
Anthropogenic alterations	dams, dredging	dams, jetties, dredging	dams, jetties, dredging
Associated River Hydrology			
Drainage area ($\times 10^3$ km^2)	24.9 (combined rivers)	4.6	13.5
Length (km)	520 (combined rivers)	210	220
Maximum elevation (m)	1,200	~500	1600
Mean Discharge (Q_w) (km^3/yr)	12.9 (combined rivers)	2.2	6.5
Total suspended sediment (Mt/yr)	0.82 (combined rivers)		0.2
Lower-River Sedimentology			
Bedload	medium sand to granules	medium sand to pebbles	medium to coarse sand
Bedforms	megaripples, sand waves, transverse bars	megaripples, sand waves	megaripples, sand waves
Terrestrial sediment sources	eskers, outwash plains, plutons	eskers, outwash plains, plutons	eskers, outwash plains, plutons
Associated Barrier System			
Barrier chain	Kennebec Barrier Chain	Saco Bay Barrier Chain	Merrimack Embayment Barrier Chain
Thickness (m)	5–10	3–11	5–20
Length (km)	11	10	21
Volume (m^3)	24×10^6	22×10^6	115×10^6
Offshore Deposits			

Glacial deposits	till, glaciomarine clay	till, glaciomarine clay	scattered drumlins and drumlin-related lag deposits, thin till cover on drumlins and bedrock, glaciomarine clay
Primary paraglacial sediment features	palaeodelta lobes at 20–30 m, 30–40 m, and 50–60 m (total volume: 2.1 x 10^9 m^3)	scattered regressive and lowstand deposits, no lowstand delta (sediment trapped in upland estuaries)	regressive braidplain delta (volume: 0.9 x 10^9 m^3), lowstand palaeodelta (volume: 1.3 x 10^9 m^3)
Holocene sediment deposits	thin (~1 m) transgressive sand and gravel deposits	thin (~1 m) transgressive sand and gravel deposits	1–9 m thick mobile sand sheet

Table 2. *Sedimentological units common to the coastal zone along river-associated paraglacial coasts of the GoM.*

Deposit	Approximate age in GoM	Relative sea-level (RSL) conditions during deposition	Sedimentology	Direct sediment source	Environment / mechanism of deposition	Associated features	Contributions to barriers
Till (non-stratified ice-contact deposits)	100–16 ka	falling sea level during glacial advance, glacial lowstand, rising RSL during deglaciation	non-sorted, non-stratified sediment with a matrix of sand and lesser amounts of silt and clay containing scattered gravel clasts and few large boulders	erosion of bedrock and pre-glacial sediments by glaciers	direct deposition by glaciers	drumlins; crag-and-tail ice-streamlined deposits; kames; ground, washboard and deglacial moraines	drumlins form pinning points for barriers, minor sediment contributions
Glaciofluvial deposits	100–16 ka	glacial lowstand, rising RSL during deglaciation	bedded gravel, sand, and mud	bedrock erosion by glaciers; erosion and reworking of glacial deposits	deposition by meltwater in terrestrial environment	eskers and outwash plains in all river basins (Fig. 3b), glaciomarine deltas, grounding-line fans	erosion of deposits in river basins provides sediment for barriers

Glaciomarine silt-clay deposits	21–13 ka	RSL rise and highstand	silty clay, fine sand, & some fine gravel, containing dropstone gravel clasts; highly compacted and dewatered; commonly sandy in the upper few meters, overlying thicker silty clay	bedrock erosion by glaciers; erosion and reworking of glacial deposits	transport to marine environment by meltwater, deposition by settling in marine environment	Presumpscot Formation in ME (Bloom 1963) , Boston Blue Clay in MA (Kaye 1961)	nearly ubiquitous deposit that form underpinnings of barriers (Figs 8, 11, 13)
Regressive shoreline deposits	17–12 ka	sea-level highstand and RSL fall	sand and fine gravel forming coastal landforms (barrier beaches, spits, regressive fluvial deltas)	erosion and reworking of sandy glacial deposits	transport to highstand / regressing shoreline by meltwater and meteoric water; reworking by waves, currents, tides and wind action along highstand and regressive shorelines	progradational deltas, beaches, spits, dunes; braidplain deltas (BPD): Sandford-Kennebunk BPD, Brunswick BPD, Merrimack BPD, parts of Kennebec River palaeodelta	deposits below modern mean sea level were partially eroded and reworked by late-Pleistocene / Holocene transgression, thus contributing coarse sediments to barrier systems

Lowstand palaeodeltas	14–12 ka	RSL fall and sea-level lowstand	fine to coarse, stratified sand and silt; bottomsets dominated by silt and clay; foresets dominated by fine, well-sorted sand and silt; topsets dominated by medium to coarse sand	erosion and reworking of sandy glacial deposits	transport to lowstand shoreline by meteoric water; reworked by waves, currents, tides and deposited as seaward-prograding bottomset, foreset and topset beds	Kennebec palaeodelta (Fig. 8); Merrimack palaeodelta (Fig. 13a); Penobscot Bay deltaic deposits	palaeodeltas partially eroded and reworked by late-Pleistocene / Holocene transgression, thus contributing coarse sediments to barrier systems
Shelf sand sheets	12 ka to present	rapidly, then slowly rising RSL	well-sorted fine to medium sand with minor quantities of silt and gravel	erosion and reworking of regressive and lowstand deposits; erosion of sandy glacial deposits in river basins	*in situ* erosion and reworking of shelf deposits; transport to lowstand shoreline by meteoric water; reworked by waves, currents and tides	mobile sand sheet in Merrimack Embayment; thin transgressive sands and gravels in Kennebec and Saco Bays	sand sheets are fraction of shelf deposits not incorporated into barriers during their formation; active exchange of shelf and barrier sediment

Estuarine deposits	8 ka to present	rapidly, then slowly rising RSL	largely massive, moderately well-sorted fine sand and silt, dominated by quartz with traces of organic material	erosion and reworking of regressive and lowstand deposits; minor bedrock and upland erosion (fluvial inputs)	onshore transport of shelf sediments; transport to backbarrier by tides through inlets and by waves as overwash across barriers	backbarrier tidal channels and tidal flats; inlet ebb- and flood-tidal deltas; common living bivalve species include *Mercenaria, Ostreidae, Ensis directus* and *Pteriomorpha*	underlie barriers; fill most accommodation behind barriers; active exchange of sediment between estuaries and barriers through inlet processes
Barrier lithosome	6 ka to present	slowly rising RSL	moderately-sorted, fine to very-coarse sand; commonly parallel laminated; coarser layers contain some granules and fine pebbles; finer layers contain very-fine sand and traces of silt.	erosion and reworking of regressive and lowstand deposits; minor bedrock and upland erosion (fluvial inputs)	onshore migration of shelf deposits; direct fluvial contributions; reworking alongshore by waves, tides and currents	barrier beaches; dunes; sandy intertidal zones; American Dunegrass common in supratidal areas	some active sediment exchange between barriers

Salt-marsh	4 ka to present	slowly rising RSL	fine-grained clastic and organic matter, fibric and hemic peat interbedded with fine sand, silt and clay; typically greater than 30% organic	*in situ* production of organic sediments; inorganic sediments largely fluvially derived	*in situ* production; inorganic sediments transported to backbarrier by tides through inlets and bydirect fluvial influx and overland flow proximal to upland areas	marsh grasses include *Spartina alterniflora* (cordgrass), *S. patens* (marsh hay), *J. gerardii* (black rush), *Phragmites* (common reed), and *Ichnocarpus frutescens* (shrub) (Jacobson & Jacobson 1989)	marsh peats underlie mainland-proximal sides of barrier complexes

Table 3. *Major rivers draining into the Gulf of Maine.*

	Location of river mouth	States / provinces in drainage basin	Drinage-basin area (km^2)
Charles River	Boston Harbor, Massachusetts (USA)	Massachusetts (USA)	1593
Merrimack River	Merrimack Embayment, Massachusetts (USA)	Massachusetts / New Hampshire (USA)	13,507
Saco River	Saco Bay, Maine (USA)	New Hampshire / Maine (USA)	4610
Androscoggin River	Popham, Maine (USA)	Maine (USA)	9376
Kennebec River	Popham, Maine (USA)	Maine (USA)	15,618
Penobscot River	Penobscot Bay, Maine (USA)	Maine (USA) / New Brunswick (Canada)	23,245
St. Croix River	Passamaquoddy Bay, Maine (USA) / New Brunswick (Canada)	Maine (USA) / New Brunswick (Canada)	3885
St. John River	St. John, New Brunswick (Canada)	New Brunswick (Canada)	7601
Annapolis River	Annapolis Basin, Nova Scotia (Canada)	Nova Scotia (Canada)	7600

Table 4. *Generalized comparison between barrier systems formed along coastal plains and those formed along, or proximal to, formerly glaciated coasts.*

	Glaciated coastal barriers	**Coastal-plain barriers**
Continuity	generally short barriers; may be sand or gravel dominated; range from barrier islands to welded barriers; barrier type commonly changes abruptly	barrier type constant for 50–200 km
Relative sea-level history during barrier formation	complex; ranges from slowly falling to slowly rising RSL during middle to late Holocene	slow rate of RSL rise during middle to late Holocene
Basement controls	drumlins and other glacial deposits and/or bedrock act as pinning points for barrier development	barriers form on interfluves, and tidal inlets stabilise in former river valleys
Sediment sources	multiple sources that can change spatially and temporally; include glacial and primary and secondary paraglacial deposits and fluvial sediments	continental shelf, minor fluvial input in locations distal to medium to large rivers
Sediment-supply rates	complex; related to fluvial and coastal erosion of glacial and paraglacial sediment sources and to RSL change	driven by sea-level change and extreme events
Substrate lithology	barrier lithosomes overlie glacial and paraglacial deposits such as till and glaciomarine clay	barrier lithosomes overlie Pleistocene coastal-plain deposits or bedrock
Grain size	fine to coarse sand and gravel; can change rapidly across short distances	fine to medium sand
Backbarrier environment	lagoon to marsh or tidal flat, incised by tidal creeks; ice-rafted horizons common in marshes	lagoon to marsh or tidal flat, incised by tidal creeks
Examples	New England (USA), Long Island (USA), Alaska (USA), Canada, New Zealand, Ireland, United Kingdom, Kamchatka Peninsula (Russia), sections of Baltic coast	East and Gulf Coasts of USA; West Africa; India; northern Black Sea; Algarve of Portugal; central and southern Brazil