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Abstract

When the Laurentide ice sheet retreated rapidly (~150 m/a) across the Penobscot Lowland between ~16 and ~15 ka, the area was isostatically depressed and became inundated by the sea. Silt and clay were deposited, but no significant moraines or deltas were formed. The Penobscot River was reborn at ~14 ka when ice retreated onto land in the upper reaches of the river's East Branch. As isostatic rebound exceeded sea level rise from melting ice, the river extended itself southward. Between ~13.4 and 12.8 ka, it established a course across marine clay and underlying glacial till in the Lowland. Its gradient was low as differential rebound had not begun. Discharge, however, was higher and the river transported and deposited outwash gravel. During the cold, dry Younger Dryas, ~11 ka, eolian sand began to accumulate in dunes in the Lowland. Some of this sand, along with fluvial sediment from the headwaters, was redistributed into terraces along gentler stretches of the river and into a paleodelta in Penobscot Bay. Eolian activity continued to ~8 ka and aggradation in terraces until ~6 ka. The climate became wetter and warmer after ~6 ka, the dunes were stabilized by vegetation, the river began to downcut, and braiding became less intense. Pauses in the downcutting are reflected in discontinuous strath terraces. In due course, the river reencountered the old outwash gravels, marine clay, glacial till, and, in a few places, bedrock. Its profile is now stepped, with gentle, gravel-bedded reaches between bedrock ribs that form rapids.

Keywords: eskers, forebulge, Holocene climate, Maine, Penobscot River, terraces

Introduction

On a map of Maine (Figure 1), there is a prominent indentation midway along the coast – Penobscot Bay – the outlet for the 175-km-long Penobscot River. With a drainage area of somewhat over 22,000 km^2 and a mean annual discharge of 340 m^3/s, the Penobscot is one of three large rivers reaching the coast of Maine.

The Penobscot heads in a series of glacially scoured lakes on an intermontane plateau, ~350 m a.s.l. (above present sea level), in Maine's western mountains. As it passes through the mountains, it descends ~250 m, entering the Penobscot Lowland near Mattawamkeag (Figure 1). Across the Lowland, the river runs largely on glacial sediment, although it has locally scoured down to greenshist-facies calcareous sandstones and impure limestones of the Siluro-Ordovician Vassalboro Formation (Osberg et al., 1985). The Native Americans had several words to describe different sections of the Penobscot River. The name Penobscot appears to be derived from *Pen apsk ek* (=descending ledges) describing the river bed, characterized by rapids, between the towns of Old Town and Bangor (Figure 1; Eckstorm, 1974: 2). After dividing around Verona Island, the river enters Penobscot Bay.

Penobscot Bay is underlain by extensively faulted late Proterozoic and Cambrian metasiltstones and interbedded metasandstones. It is bordered on the west by Ordovician schists and pelites and on the east by Precambrian and Cambrian volcanoclastics, and Devonian and Silurian volcanics and granites (Osberg et al., 1985). The dextral Penobscot Bay and Turtle Head fault zones run the length of the bay, parallel to and on either side of Islesboro, a 27-km-long string of islands (Figure 1). The bay appears to owe its existence to erosion along these fault zones and along a number of smaller en echelon normal faults running diagonally between them (Osberg et al., 1985; Stewart, 1998). Scoured into the bedrock on either side of Islesboro are troughs, each containing overdeepened depressions that were presumably largely glacially eroded (Knebel and Scanlon, 1985).

The topography of this area—the western mountains, the Penobscot lowland, and Penobscot Bay—is a product of nearly 400 m yr of fluvial erosion of a Rocky-Mountain scale Devonian landscape from which an average of nearly 10 km of rock has been striped (Hooke and Winski, 2014). During the last million years or so, a succession of glacial advances have modified the fluvial landscape.

Herein, we focus on a section of the river in the Penobscot Lowland where the changes in climate during the Holocene led

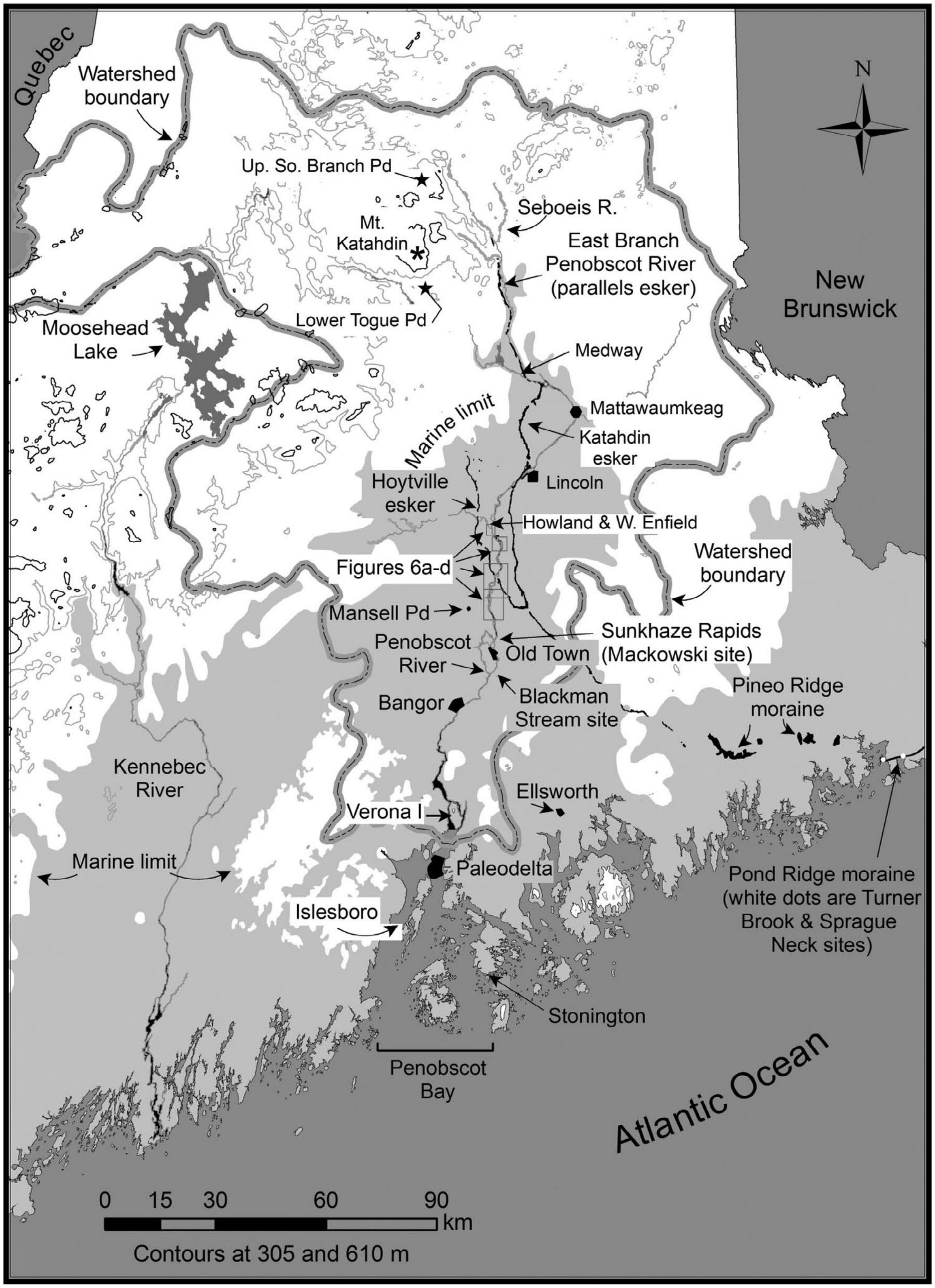

Figure 1. Map of Maine showing the Penobscot River, its drainage basin, and locations mentioned in the text. Light gray areas on land were submerged during ice retreat. Light contour is at 305 m and heavier one is at 610 m a.s.l. Marine limit and Pineo Ridge moraine from Thompson and Borns (1985).

to the formation of eolian dunes and river terraces. First, however, we describe our methodology, the likely effect of isostatic rebound on the river, and the stratigraphy. A companion paper (Hooke and Hanson, in preparation) deals with the East Branch of the river north of Medway (Figure 1).

Methodology

Quaternary units bordering the Penobscot River in the Lowland were initially mapped by Lowell (1980), Borns (1981), Borns and Thompson (1981), Smith and Thompson (1986), and Newman (1986a, 1986b) at a scale of 1:62,500. These materials have been studied in greater detail over the past decade by Hooke (in preparation), Hooke et al. (2016), Kelley and Kelley (2012), Hildreth (2011), and Syverson and Thompson (2011).

Our identifications of map units are based on topography and sedimentology. Vertical exposures being rare, sedimentology was normally studied by sampling with an 18-mm soil probe thrust into the surface to a depth of 0.1–0.9 m, depending on the grain size and stiffness of the material. Sand sizes were evaluated by comparison with images on an AMSTRAT transparent sediment-size card viewed with a 10× hand lens. A few samples were collected for analysis in a settling tube.

The primary topographic landforms are gravel ridges (eskers), dunes, and relatively planar surfaces identified either as parts of the present floodplain or as braidplains forming terraces above the present floodplain. The terraces are rarely continuous enough to correlate simply by projecting the terrace surface across gaps. Rather, we used elevation above the modern floodplain as the primary tool for making correlations. We chose to define the floodplain as the level of the mean annual flood, the flow with a recurrence interval of ~2.33 years (Pizzuto, 1986). We determined the gage height of this flood, 4.6 m, from records of annual peak discharge at the USGS West Enfield gaging station (Figure 1; see Leopold et al., 1964: 63 for technique). This record spanned 99 years at the time of the calculation.

The heights of terraces above river level were initially determined with a hand level, and later by surveys with a Nikon AE-5 automatic level. The latter surveys were anchored on USGS 'recoverable marks', which are commonly elevations on relatively level sections of roads. Profiles were generally "closed" by returning to a known elevation, such as a second recoverable mark or an approximately level water surface (such as the river between riffles).

On each day that we determined such an elevation, we obtained the gage height at West Enfield and thus estimated the height of the floodplain above the river level on that day at the time of the survey. Recognizing that there is relief of ~±1 m on the floodplain of a river the size of the Penobscot, we mapped surfaces that were below and up to about a meter above the mean annual flood level as floodplain; higher surfaces were mapped as terraces. Only one terrace was continuous enough to suggest a common time of formation, and thus to warrant a name—the Sugar terrace. Others are inferred to be strath terraces formed during temporary and local pauses in downcutting from the Sugar level and are referred to by their height above the modern floodplain.

Samples of sand from eolian dunes, river terraces, and paleochannels were collected for OSL dating. About half the samples were obtained by auguring; a 1½″ × 6″ PVC tube was pressed into the sand in the auger head when it was raised from the desired sampling depth. The remaining samples were collected from steep faces by digging back several decimeters to expose undisturbed stratigraphy and hammering 2″ × 7″ PVC tubes into the resulting face. After removing a tube from the auger head or face, excess sand was trimmed leaving a sand surface flush with the tube ends to minimize disturbance during shipping, and the ends were covered with multiple layers of duct tape. Most samples were collected from depths >1 m below the present ground surface (Table 1) to minimize potential problems associated with bioturbation and post-burial disturbance (Bateman et al., 2003; Hanson et al., 2015).

OSL samples were processed at the Luminescence Geochronology Laboratory at the University of Nebraska–Lincoln. OSL measurements were made on a Riso model DA-20 luminescence reader, and equivalent dose (D_e) values were calculated using the single-aliquot regenerative (SAR) method (Murray and Wintle, 2000). A preheat plateau test (Wintle and Murray, 2006) was used to determine the optimum preheat temperature. Aliquots were rejected if their recycling ratios were >10% or if they had measurable signals during stimulation with IR diodes. A minimum of 20 accepted aliquots were generated for each sample, and final age estimates were calculated using the Central Age Model (Galbraith et al., 1999). Environmental dose rate estimates for the OSL samples were based on elemental concentrations of bulk sediment taken from within ~30 cm of the OSL sample. These bulk samples were analyzed for concentrations of K, U, and Th using a high-resolution gamma spectrometer. The cosmogenic component of the dose rate values was calculated using equations from Prescott and Hutton (1994), and the final dose rates were calculated following equations from Aitken (1998). Age estimates, dose rate values, and equivalent dose values are shown in Table 1. In three of the dunes and one of the terraces, samples were collected at different depths; ages were in correct stratigraphic order in two cases (Table 1, Dates #9 and 10, and #18 and 19) and were the same age within 1σ limits of uncertainty in the other two (Table 1, Dates #3 and 4, and #5 and 6).

Five samples were collected for ^{14}C dating. One was a *Hyatella arctica* marine bivalve shell from an exposure of glaciomarine mud in an esker (see SOL), two were bulk samples of peat, and two were charcoal or wood fragments from peat. The peat samples were obtained from depths of 1–2. m with a Dutch Corer (Eijkelkamp Company, The Netherlands). One of the bulk peat samples proved to be contaminated by modern rootlets and did not yield a useful age. All ^{14}C ages were calibrated using CALIB 7.0 (Stuiver et al., 2014). For most of the dates on marine material, we followed Borns et al. (2004: 91) and used a 600-year marine reservoir correction. However, the marine reservoir correction may vary both spatially and temporally, and for dates associated with the Pond Ridge and Pineo Ridge moraines mentioned later we used the standard 400-year correction to achieve better agreement with the North Greenland Ice Core Project (NGRIP) ice core $\delta^{18}O$ record (see below).

Isostatic adjustment

The load of an ice sheet depresses Earth's crust. This causes an upward deflection of the crust, called a forebulge, 150–200 km in front of the ice sheet (Hooke and Ridge, in press; Figure 3). Under appropriate conditions, this forebulge migrates northward as the ice margin retreats, affecting river gradients.

Balco et al. (1998) showed that owing to isostatic depression, Moosehead Lake (Figure 1) drained northward into the Penobscot River immediately after deglaciation. Radiocarbon dates from basal sediment filling the outlet channel suggest that this northern outlet was abandoned at ~9.8 ± 0.3 ka (Kelley et al., 2011; Figure 3). Abandonment is attributed to passage of the forebulge beneath the lake, tilting the land to the south. The tilting raised the northern outlet above the present one, which is ~30 km further south and delivers water to the Kennebec River. Later, the land tilted slightly back to the north. Balco et al. (1998; Figure 2) place the peak of the forebulge under the lake at ~9.5 ka and the subsequent trough sometime between 7.8 and 5.7 ka.

A well-established relative sea-level curve (sea level relative to present sea level: long dashed line in Figure 2) suggests that a forebulge may have migrated across the coast at ~12.2 ka. The land elevation, relative to sea level (solid line in Figure 2), can be estimated by subtracting relative sea level from global sea level (e.g. Barnhardt et al., 1995). There is a distinct high in the land elevation at ~12.2 ka. A problem with interpreting this as passage of a forebulge is that the duration of the period of high land elevation, ~1 ka, is far too short to correspond to the passage of a bulge with a likely wavelength of >200–300 km (e.g. Brotchie and Silvester, 1969; Turcotte and Schubert, 2002, equations (3)–(130)),

Table 1. Dates.

Context	Sample number	Depth, m	Age ± 1σ (ka)	Laboratory Number	Location Latitude	Location Longitude	U (ppm)	Th (ppm)	K_2O (wt %)	In situ H_2O (%)[a]	Dose rate (Gy/ka)	CAM[b] De (Gy) ±1 Std Err	Aliquots (n)[c]	OD[d] %	^{14}C age ±1σ
Esker															
1. Shells dragged(?) up onto esker ^{14}C (Figure 6c)[e]	Greenbush-1		13.8f,g ± 0.1	GX-29787-AMS	45.10961	−68.62639									12,520 ± 50
River gravels															
2. Old gravel beneath modern floodplain (Figure 6b)	4673	2.2	11.4 ± 1.2	UNL-3113	45.20353	−68.62100	1.9	6.6	2.8	0.6	3.31 ± 0.19	37.6 ± 2.8	25/50	36.8	
Dunes															
3. Easternmost dune (top) (Figure 6d)	3557t	1.0	10.8 ± 1.1	UNL-2132	45.04151	−68.62493	1.8	7.5	2.5	6.4	2.93 ± 0.21	31.6 ± 1.9	27/40	29.6	
4. (bot)	3557b	3.0	10.6 ± 1.4	UNL-2133	"	"	1.6	7.1	2.1	6.3	2.52 ± 0.18	26.7 ± 2.6	25/40	46.9	
5. South of Sugar Island (top) (Figure 6c and d)	3513t	1.6	9.9 ± 1.1	UNL-1884	45.09177	−68.64957	1.6	6.9	2.5	5.9	2.86 ± 0.20	27.4 ± 2.0	23/42	32.3	
6. (bot)	3513b	2.6	8.7 ± 0.9	UNL-1885	"	"	1.6	7.0	2.5	3.4	2.93 ± 0.18	25.6 ± 1.8	23/33	31.9	
7. Dune west of Penobscot River (6c)	5052	1.1	9.1 ± 0.9	UNL-3481	45.14828	−68.64051	2.1	7.8	2.9	5.7	3.33 ± 0.23	30.2 ± 1.8	34/51	34.2	
8. Dune blocking E paleochannel (Figure 6c and d)	5050	0.9	8.9 ± 1.0	UNL-3480	45.06906	−68.63044	1.7	7.1	2.6	6.8	2.97 ± 0.22	26.3 ± 1.8	28/36	36.8	
9. Dune 0.7 km south of SE corner of Figure 6d (top)	3556t	1.1	8.8 ± 0.9	UNL-2130	44.99390	−68.62708	1.7	6.5	2.8	3.3	3.20 ± 0.20	28.0 ± 1.9	20/40	27.1	
10. (bot)	3556b	2.8	9.4 ± 0.9	UNL-2131	"	"	1.5	5.4	2.8	4.8	2.93 ± 0.20	27.5 ± 1.6	23/40	25.5	
11. Dune 3 km NNW of Howland (N of Figure 6a)	5054	1.3	8.3 ± 0.8	UNL-3482	45.26467	−68.67305	1.6	6.5	3.1	6.0	3.25 ± 0.24	27.1 ± 1.4	29/42	25.6	
12. Dune blocking middle paleochannel (Figure 6c and d)	5046	0.9	7.7 ± 0.8	UNL-3479	45.07684	−68.63945	1.8	6.9	2.8	5.1	3.19 ± 0.22	24.5 ± 1.8	34/50	40.5	
Paleochannel filling															
13. Peat from bottom of western paleochannel (Figure 6c and d)	5040 (GR-1)	1.4	9.5 ± 0.1	Beta 310404	44.97959	−68.64933									8460 ± 50
14. Sand from just beneath fill in western channel (Figure 6c and d)	5802	1.3h	7.5 ± 2.4	UNL-4016	"	"	1.9	6.0	3.0	38.4	2.37 ± 0.71	17.8 ± 1.6	38/55	52.5	
15. Peat from bottom of eastern paleochannel	5168	2.5	Contaminated	Beta 333423											730 ± 30
16. Wood from bottom of Passadumkeag paleochannel (Figure 6b)	5695-4	1.1	4.1 ± 0.0	Beta 398252	45.20735	−68.62268									3760 ± 30
17. Charcoal from bottom of eastern paleochannel (Figure 6d)	5720	2.4	3.3 ± 0.0	Beta 398251	45.06152	−68.62766									3070 ± 30
Terraces															
18. 8.9 m terrace in Greenbush (top) (Figure 6c and d)	3687t	1.2	6.7 ± 0.7	UNL-2353	45.08740	−68.65011	2.1	7.5	2.9	6.2	3.32 ± 0.24	22.1 ± 1.3	23/34	23.3	
19. (bot)	3687b	3.6	8.1 ± 0.9	UNL-2354	"	"	1.9	6.2	2.9	6.5	3.13 ± 0.24	25.5 ± 1.8	24/39	32.5	
20. Sugar terrace in Greenbush locality (Figure 6c and d)	4120b	2.8	7.9 ± 1.1	UNL-2637	45.08093	−68.65011	4.9	9.1	2.6	15.3	3.46 ± 0.42	27.4 ± 1.2	22/44	18.7	
21. Sugar(?) terrace on Olamon Island (Figure 6c)	4118b	2.5	7.0 ± 0.6	UNL-2636	45.21654	−68.62801	4.0	8.1	3.1	2.1	4.09 ± 0.23	28.7 ± 1.7	20/29	25.9	
22. 2 m terrace near S edge of Figure 6d	4661	1.0	Not datable	UNL-3111	45.09669	−68.65090	–								
23. 1 m terrace in Passadumkeag locality (Figure 6b)	4663	1.1	9.0 ± 0.8	UNL-3112	45.20353	−68.62100	1.7	6.6	3.1	0.3	3.53 ± 0.20	31.7 ± 1.9	35/50	33.5	
24. 1 m terrace in Passadumkeag locality (Figure 6b)	4674	0.3	7.5 ± 0.6	UNL-3114	45.20977	−68.62994	1.8	6.7	2.8	0.6	3.34 ± 0.19	25.1 ± 1.1	21/25	19.2	

a. Assumes 100% long-term variability in measured moisture content.
b. Central Age Model (Galbraith et al., 1999).
c. Accepted disks/all disks.
d. Overdispersion.
e. Number of figure on which site is located.
f. Reservoir correction of 600 years assumed.
g. See Supplementary Material, available online.
h. Location ~8 m east of sample 5040, where peat was ~0.2 m thinner.

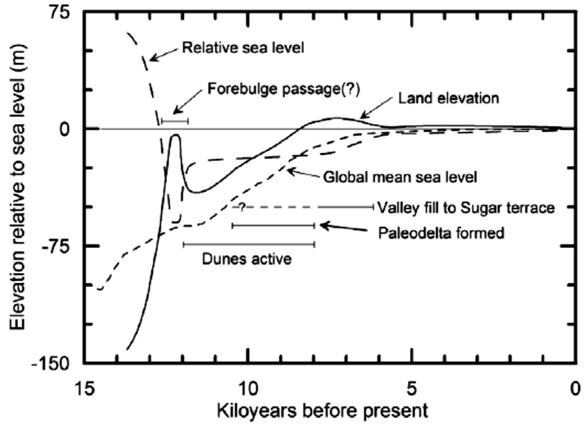

Figure 2. Relative sea level curve for coastal Maine (Kelley et al., 2013) and estimate of land elevation obtained by subtracting relative sea level from global sea level (Lambeck et al., 2014). (Compare with Barnhardt et al., 1995).

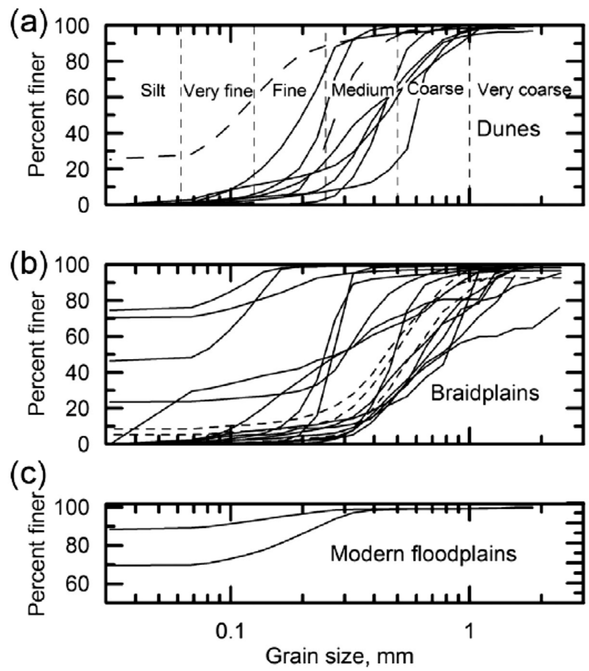

Figure 3. Grain size analyses, by settling tube, of samples from (a) features mapped as dunes, (b) features mapped as braidplains, and (c) features mapped as modern floodplains. In (a), the two distributions shown with long dashed lines are samples from the auger hole that yielded the OSL date of 7.7 ka, and in (b), the three distributions shown with short dashes are from the terrace section that yielded an OSL age of 7.5 ka.

at a reasonable speed. Barnhardt et al. (1995), citing Clark et al. (1994), suggest that local inhomogeneities in the crust may have modified the crustal response—a distinct possibility considering the presence of Mesozoic rift basins offshore and parallel to the coast (Ballard and Uchupi, 1972).

If the forebulge passed beneath the coast at ~12.2 ka and beneath Moosehead Lake at ~9.5 ka, the migration rate was ~67 m/a.

The ice sheet margin was at the edge of the continental shelf at the Late Glacial Maximum (LGM) stand, ~23 ka. If the peak of the forebulge was then ~200 km outside the margin and passed beneath Moosehead Lake at ~9.5 ka, a distance of ~800 km, the migration rate was ~60 m/a. These rates are broadly consistent with each other and with the mean rates of ice margin retreat: ~60 m/a from the shelf edge to the coast and ~90 m/a from the coast to Moosehead Lake. We thus think it likely that the gradient of the Penobscot was affected by a migrating forebulge during the Holocene.

Influence of isostatic rebound and of a migrating forebulge on the Penobscot River

Analysis of the potential influence of isostatic rebound and of a migrating forebulge on the river starts with consideration of isostatic adjustment in the broader New England region, and in particular of paleoshoreline features formed in several glacial lakes in western New England. These features define planes that slope upward to the north at ~0.9 ± 0.1 m/km (Koteff and Larsen, 1989). The azimuth of the maximum inclination in the Connecticut River Valley, the site of Glacial Lake Hitchcock, is ~N20W (Koteff and Larsen, 1989: 113), but it is not well constrained. The regional uniformity of the inclination, or tilt, implies that it is likely related to fundamental properties of the crust such as its mechanical thickness and elastic modulus.

Shoreline features formed by Moosehead Lake at 11.5 ka are now tilted upward to the N60W at an apparent slope of only 0.7 m/ km (Balco et al., 1998). Older shoreline features have a ~1 m/ km tilt, but the timing is unknown.

Numerical modeling (Hooke and Ridge, in press) suggests that following deglaciation, isostatic rebound is spatially uniform for a few millennia before differential rebound and tilting begin. Deglaciation of the Connecticut River valley began ~17.9 ka and was largely complete by 13.9 ka, but differential rebound apparently did not start until ~14.5 ka. The rate of rebound decreases exponentially with time. Differential rebound at Moosehead almost certainly started before 11.5 ka; if it started at, say, ~12.3 ka, the initial northward depression of the landscape would have been ~0.9 m/km, consistent with observations in western New England.

The present mean gradient of the Penobscot River from Medway to Sunkhaze Rapids (Figure 1) is ~0.4 m/km. This section of the river is at about the same latitude as the glacial lakes farther west in New England and was deglaciated at about the same time. There are large bogs in the Lowland that may once have been shallow lakes, but there is no known evidence for deeper postglacial lakes. Thus, the Penobscot likely had a (small) positive slope toward the coast when it was re-established across the Lowland at ~13.5 ka. Making an adjustment for a ~1 ka time lag between the initiation of differential rebound at ~14.5 ka and emergence of the Lowland above sea level at ~13.5 ka, the maximum northward depression of the landscape along the river's course must have been less than ~0.5 m/km.

The azimuth of maximum tilt in the Lowland likely lies between that of Moosehead and Lake Hitchcock, but much closer to that of Moosehead. The component parallel to the Penobscot of a 0.9 m/km tilt to the N50W, for example, is 0.5 m/km.

The implications of these calculations are as follows: (1) a regional tilt of 0.9 m/km with azimuth varying eastward from western New England is consistent with the available data, (2) the Penobscot likely had a near-negligible gradient when it was reestablished across the Lowland, and (3) its gradient has been steadily increasing since that time. As the forebulge passed beneath the Lowland, the gradient would have increased slightly and then decreased again before resuming its steady increase to today's value.

Stratigraphy

Eskers and associated material

The dominant facies in eskers is gravel consisting of subrounded pebbles and small cobbles supported by a matrix of coarse sand. Large cobbles and small boulders make up a minority of the clast sizes. At the very base of an esker core, however, beds of rounded boulders ≥0.4 m in diameter and ordinarily a single particle diameter in thickness are occasionally seen or reported by gravel pit operators.

Eskers are formed where heat dissipated by subglacial streams melts conduits upward into the ice, releasing sediment entrained in the basal ice and overloading the stream (Shreve, 1972). Under thick ice, negative feedbacks lead to a rough balance between melting of the conduit walls and closure of the tunnel by ice flow (Hooke, 2005: 232–241). Near an ice sheet margin, however, the ice is too thin to close a tunnel as rapidly as melting enlarges it. The resulting enlargement of the tunnel decreases the water velocity, leading to more rapid deposition. During periods of still stand or slow margin retreat, quite massive deposits may accumulate here, forming *esker heads* (Hooke and Fastook, 2007).

Commonly found downflow from Maine esker heads are beds of salt-and-pepper gray (5Y6/1) medium quartz-ofeldspathic sand with 10–15% mafic grains. We think these sands are bed-material load deposited either in such an enlarged tunnel near the ice margin or in the proglacial environment, relatively close to the ice margin.

Also frequently observed are beds of tan (5YR6/4) fine quartz-ofeldspathic sand. These may either be interbedded with the gray medium sand in millimeter to centimeter-scale laminations or more commonly draped over the esker anywhere along it. The fine sand is likely wash load, carried in suspension and thus widely distributed away from the ice margin. It is basically a nearshore equivalent of the widespread Presumpscot Formation, a locally fossiliferous marine mud (Bloom, 1960, 1963; Thompson and Borns, 1985).

Dunes

Dunes were identified by their topography and by their typical composition of moderately well-sorted, medium sand (Figure 3a). Distinctive dune topography is not always present or is obscured by dense vegetation, and in these cases identification is based on composition alone. Stratification is rarely exposed.

Penobscot River floodplain and braidplain sediment

The fluvial sediment of concern is silt and sand deposited in floodplains and braidplains. In the braidplains, the sediment ranges from fine sand to fine gravel and is generally poorly sorted (Figure 3b). Bedding on a centimeter to decimeter scale is common. Similarity of grain size and geomorphic relations suggest that much of the sand in braidplains is remobilized from dunes with some winnowing of the finer grains. The modern floodplain is typically composed of poorly sorted silt and very fine sand (Figure 3c).

Deglaciation and eskers

During the maximum phase of the LGM, ~20,000–22,000 years ago (Lambeck et al., 2014), ice extended across the Gulf of Maine to the edge of the continental shelf, and Earth's crust was depressed 300–400 m below present sea level (Hooke and Fastook, 2007; Figure 3). By ~16 ka, the ice margin had retreated to roughly the present coastline and, although substantial rebound had occurred, shoreline features presently at ~73 m a.s.l. in the vicinity of Penobscot Bay (Belknap et al., 1987; Thompson et al., 1983; Figure 1) demonstrate that the present coast was still submerged. With further ice retreat, the sea advanced inland, submerging the Penobscot Lowland and eventually extending some distance up into the foothills of the western mountains (see "Marine limit" in Figure 1). By ~14 ka, the rate of isostatic rebound began to exceed that of sea-level rise, and the land started to emerge.

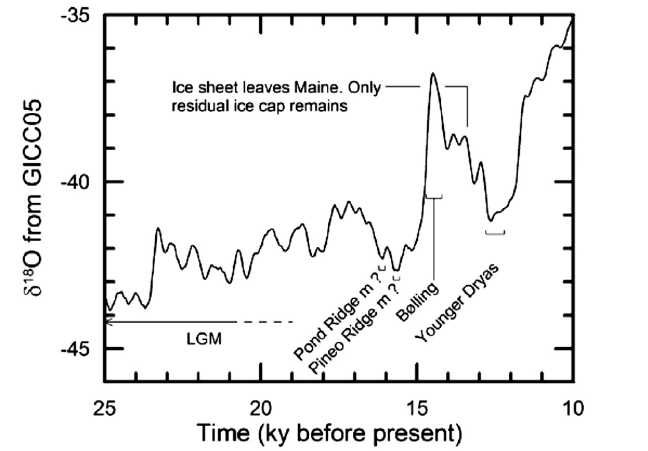

Figure 4. Smoothed $\delta^{18}O$ record from ice cores from Greenland (Andersen et al., 2006). More negative $\delta^{18}O$ values are indicative of colder temperatures.

As the marine ice margin retreated inland, vigorous streams emerged from beneath the ice and built submarine fans along the margin (Ashley et al., 1991; Borns et al., 2004; Dorion et al., 2001; Hooke and Fastook, 2007; Hunter et al., 1996). Where the margin was stable for a sufficiently long period of time, the fans reached sea level and became deltas.

Eskers formed within 5–10 km of the margin. Farther inland, temperature gradients in the basal ice were apparently high enough to conduct all the heat dissipated by the subglacial streams upward into the ice (Hooke and Fastook, 2007). Thus, the eskers were formed in segments as the ice retreated. The segments are aligned with one another to form discontinuous ridges, tens of kilometers in length. One, the Katahdin esker (Stone, 1899: 104–115), lies subparallel to the Penobscot River over much of its course (Figure 1).

Dating the retreat

Two ^{14}C dates (Dorion et al., 2001) on marine shells interbedded with diamicton on the ice-distal side of the Pond Ridge moraine at Turner Brook (Figure 1) yield a calibrated mean age of 16.0 ± 0.2 ka. Consistent with this age are (1) a minimum age of 15.5 ± 0.2 ka from glaciomarine facies overlying the moraine at Sprague Neck, 5 km west of Turner Brook (Dorion et al., 2001) and (2) a maximum age of 16.4 ± 0.1 ka from laminated silt underlying diamicton at Dennison Point, 3 km outside the moraine (Kaplan, 1999). Additional ^{14}C dates (Dorion et al., 2001) from ice-proximal facies in a core from Lily Lake and from basal glaciomarine facies in a core from Marks Lake, both of which lie just north of the Pineo Ridge moraine (Figure 1), yield mean *minimum* ages for that moraine of 15.5 ± 0.1 and 15.4 ± 0.1 ka, respectively. Oxygen isotope minima in the NGRIP ice core occur at 16.1 and 15.7 ka (Figure 4; GICC05 chronology from Andersen et al., 2006). Although the $\delta^{18}O$ record may reflect largely winter temperatures, particularly when the changes are large (Denton et al., 2005), we think it still worth considering the possibility that these minor cold events resulted in pauses in the retreat, or possibly slight readvances, and are thus responsible for these moraines. Hooke and Fastook (2007) suggested this correlation, based in part on distinct pauses visible in a computer animation of the retreat, but used a less refined time scale.

Davis et al. (2015) argued that Mt. Katahdin in northern Maine (Figure 1) was a nunatak at this time, and that moraines on the eastern and southern sides of the mountain were formed by continental ice that surrounded the mountain. They obtained

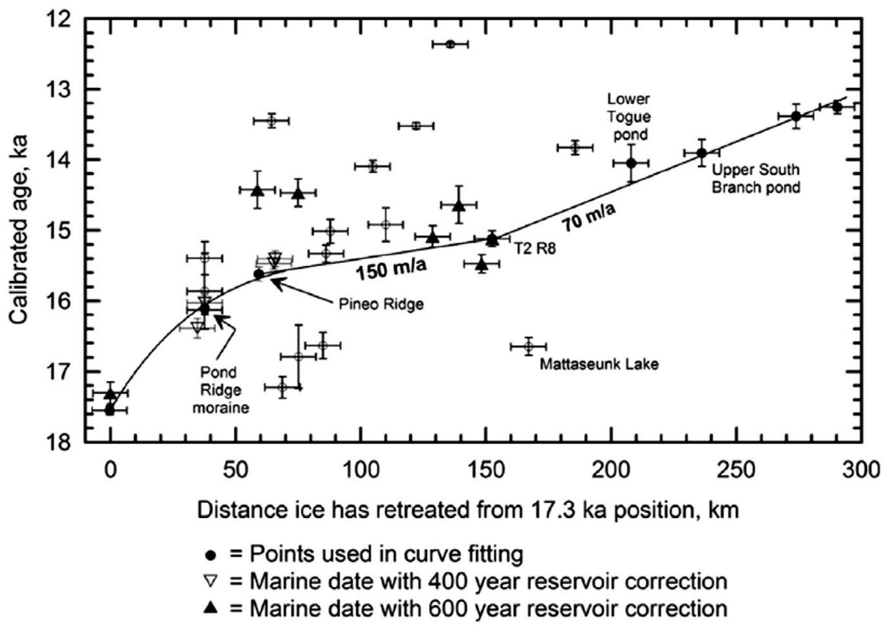

Figure 5. Time–distance plot of ice sheet retreat. Points are from sites reported by Borns et al. (2004) lying within ~50 km of a transect running from the coast, through Ellsworth, and ~10 km west of Lower Tongue Pond (Figure 1), roughly normal to Borns et al.'s contours of deglaciation age, supplemented with others mentioned in the text. Curve is based on seven points shown with •. Much of the scatter is likely because of the errors in interpretation of the context of the dates and to lateral (parallel to the ice front) differences in the rate of retreat.

a ^{10}Be exposure age of 16.1 ± 1.2 ka on the Basin Ponds moraine on the eastern side of the mountain and suggested that it might be correlative with the Abol moraines on the south side of the mountain and with the Pineo Ridge moraine. We propose, instead, that it and the upper of two prominent Abol moraines may be correlative with the Pond Ridge moraine, and the lower of the Abol moraines with the Pineo Ridge moraine. Such a correlation would be consistent with the fact that a small amount of thinning (the Abol moraines are ~50 m apart vertically) commonly leads to a much larger margin retreat (~9 km between the Pond Ridge and Pineo Ridge moraines) and is consistent with known radiometric dates.

Near Lincoln (Figure 1), a ^{14}C date on a mollusk from near the base of some glaciomarine mud (Dorion et al., 2001, Site T2 R8 NWP) yielded a calibrated age of 15.1 ka (Figure 5). Dorion et al. (2001) interpret this as a "close minimum" age for deglaciation (p. 235). This suggests a retreat from the Pineo Ridge moraine to this point, a distance of ~90 km, in ~600 years, or 150 m/yr. This is consistent with the spacing of small moraines in the Penobscot Lowland within ~100 km of the coast (Eusden, 2014). The moraines are unmistakable on Lidar images and are likely nearly annual. The spacing ranges from 60 to 280 m depending, in part, on slope of the bed.

On the other hand, Dorion (1997) obtained samples from marine silts resting on a diamicton, interpreted to be lodgement till, in Mattaseunk Lake, about 11 km east of Medway. From a horizon 1.8 m above the diamicton, the periostraca (a part of the organism commonly composed of terrestrial, not marine C) of *Portlandia arctica* produced a ^{14}C age of 13,450, and terrestrial vegetation from the same horizon, a ^{14}C age of 13,500 (Jenkins and Dorion, 2013: 143). These yield a mean calibrated age of 16.2 ± 0.1 ka. The marine sediment beneath the horizon contained 501 sediment couplets. If these are annual, the top of the diamicton dates to ~16.7 ka. This date appears robust, but it is clearly inconsistent with others discussed herein (Figure 5).

Further north, dates from Lower Tongue and Upper South Branch ponds suggest that the retreat rate slowed to ~70 m/a (Figure 5; Hooke and Hanson, in preparation).

Early post-glacial history

The Penobscot River was reborn at ~14 ka when, in the valley of the East Branch, the margin retreated onto land near the present- day confluence with the Seboeis River (Figure 1; Hooke and Hanson, in preparation). The Penobscot extended itself headward as the ice sheet retreated and southward as isostatic rebound and relaxation of the gravitational pull of the ice sheet on the sea (e.g. Clark, 1976a, 1976b) outpaced global sea level rise from melting ice sheets. Relative sea level fell precipitously (Figure 2). Rough scaling suggests that the river was establishing its course across the Penobscot Lowland between ~13.4 and 12.8 ka. The river gradient was likely low, as differential rebound probably had not yet started. Discharges, however, were likely higher at this time as the melting ice sheet was still contributing water and Moosehead Lake was still draining into the Penobscot.

A bed of matrix-supported pebbles and small cobbles is commonly present at river level and in lower terraces along the East Branch and southward to the coast. The gravel is typically overlain by 1.5–2 m of fining upward, silty, fine to medium sand. In the Passadumkeag area, this gravel is widely exposed in shallow quarries (Figure 6b). Locally, it appears to overlie cobble and boulder gravel. Some of the boulders, particularly the larger ones, are likely lag produced from erosion of till. At one locality, at a depth of ~2 m beneath the modern floodplain, a bed of medium to coarse sand interbedded with the gravel yielded an OSL age of 11.4 ± 1.2 ka. At a depth of ~1 m in this same exposure, there is a transition to medium and then to fine sand. An OSL date of 9.0 ± 0.8 ka on well-sorted coarse sand at this transition suggests that deposition of gravel may have continued here until somewhat before ~9.0 ka.

Still farther south, at the Mackowski archeological site, just above Sunkhaze Rapids (Figure 1), a hawthorn seed from a depth of ≥~1 m was dated to 9.5 ka (Kelley and Sanger, 2014; Robinson, 2006). The seed was underlain by 1.8 m of silty sand and sandy silt, which, in turn, rested on cobble gravel.

At another archeological site, the Blackman Stream site adjacent to the Penobscot River (Figure 1), a Bsb buried soil horizon

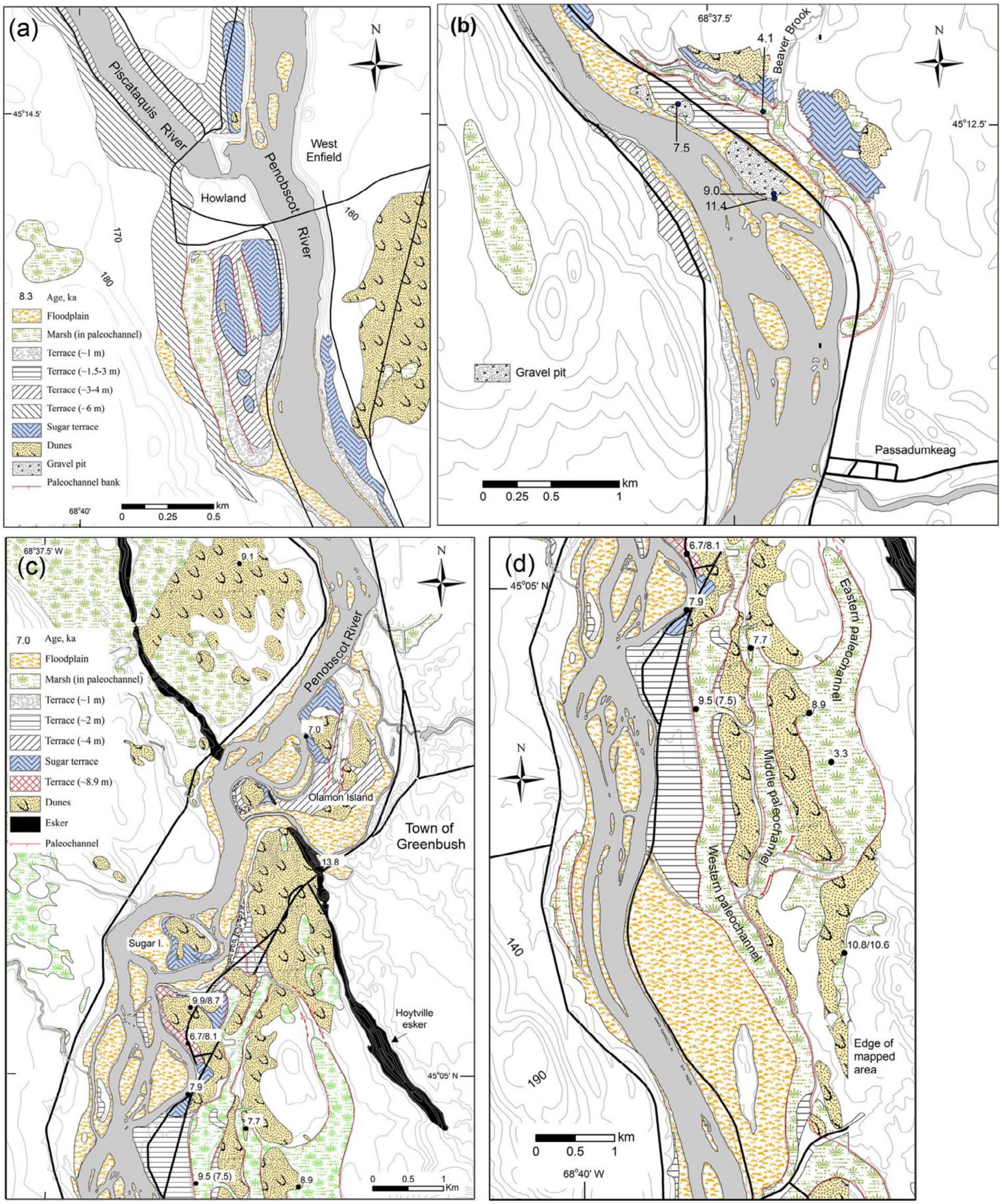

Figure 6. Maps of relevant Quaternary units. Blank areas in all panels are underlain by materials not relevant to present discussion, commonly either glacial till or glaciomarine Presumpscot Formation. For explanation of symbols, see panels (a) or (c). (a) Detailed map of dunes and terraces in northern Howland quadrangle; (b) detailed map of dunes and terraces in Passadumkeag; (c) detailed map of dunes and terraces in the Greenbush area; and (d) detailed map of dunes, terraces, and paleochannels south of Olamon Island.

at a depth of ~1 m, contained charcoal. Three samples were dated, yielding ages of 8.2, 8.6, and 9.3 ka (Kelley and Sanger, 2014; Sanger et al., 1992). The soil is underlain by ~1 m of fine sand and silt which, in turn, overlies stratified gravel, sand, and silt. A projectile point just above the gravel is of a style that is commonly associated with occupations dating from ~10 ka.

Finally, near the head of Penobscot Bay, Knebel and Scanlon (1985: 320) cite reports of a bed, <2 m thick, of pebbly and silty sand or sandy gravel overlain by finer sediment. From this coarser unit, two *Mya arenaria* in life position yielded identical ages of 9.2 ± 0.1 ka (Barnhardt et al., 1997).

We suspect that the gravels at all these sites are part of a continuous stratum deposited as the reborn Penobscot River was extending itself southward and while there was still melting ice in northern Maine feeding higher discharges. The dates suggest that the deposition of gravel continued until ~9–10 ka.

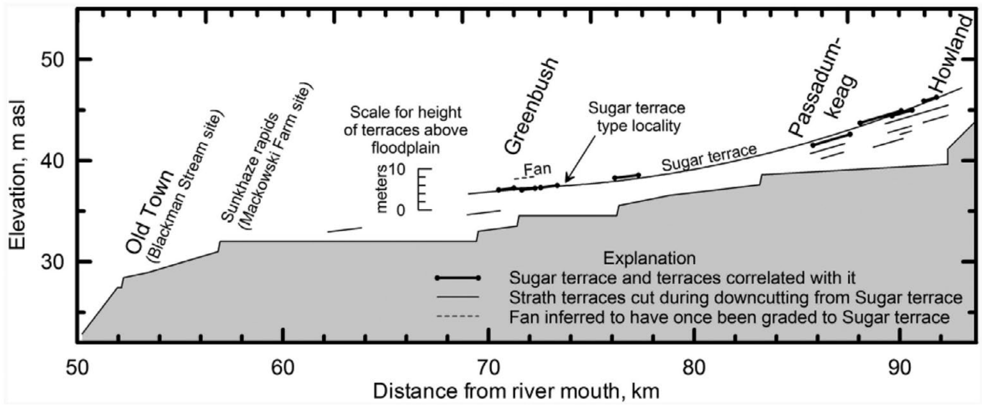

Figure 7. Longitudinal profile of Penobscot River from Howland to slightly south of Old Town, showing terraces.

Transition to fine-grained sedimentation

At all the sites mentioned above, there is an abrupt transition from gravels to fine sand and silt at a depth of 1–3 m. The decrease in runoff as residual ice melted (~11 ka; Borns et al., 2004) and the ~15% decrease in discharge when Moosehead Lake began to drain to the Kennebec at ~9.8 ka (Kelley et al., 2011) are likely largely responsible for this transition. The competence of the river was also decreased where erosion exposed transverse bedrock sills in the bed, reducing the gradient upstream from them (Kelley and Sanger, 2003: 125).

This was a time period, 12.2–9.5 ka, during which we suspect that a forebulge was migrating northward through this section of the river, first steepening the gradient in its lee, and then reducing it to something approaching its present value. As some differential rebound was likely occurring, there was probably also a progressive increase in gradient through time on the proximal side of the migrating forebulge.

The remainder of this paper concerns the period from ~12 to ~6 ka and focuses on the section of the river extending from Howland to somewhat north of Old Town (Figures 1 and 6a–d).

Dunes

Sometime after rebirth of the Penobscot River, and after sea level had fallen to roughly its present level (~12.7 ka) on its way to −60 m, eolian sand dunes began to accumulate on both sides of the river (Figure 6a–d). Ten OSL dates on sand from these dunes fall between 10.8 ± 1.1 and 7.7 ± 0.8 ka (Table 1). The dates provide an estimate of the time when the sand at the sample depth was last actively blown about, and thus subjected to solar radiation. As the samples were from depths of <3 m, the actual accumulation of the sand started earlier than the dates obtained.

We suspect the dunes began to form prior to and during the Younger Dryas (12.7–11.6 ka) under periglacial conditions. An exhaustive analysis of pollen data available prior to ~1992 suggests that the climate in Maine from ~13 to at least ~7 ka was cooler and *significantly drier* than at present (Webb et al., 1993: 451–453). By 3 ka, precipitation had increased to approximately present levels, but July temperatures were warmer than today. This pattern is consistent with data from Mansell Pond, ~5 km west of the Penobscot River (Figure 1; Almquist et al., 2001). In the deepest core from Mansell Pond, a ^{14}C date from gyttja immediately above refusal in the nearly impermeable marine Presumpscot Formation yielded an age of 10.9 ka, suggesting that prior to that time the water balance (precipitation minus evaporation) was not sufficient to maintain a lake. Between 9 and 4 ka, the lake level was more than 6 m below its present level and peat was being formed.

In western Maine, east of the Kennebec River, Borns and Hagar (1965) and McKeon (1989) also mapped a sand blanket and dunes with grain sizes similar to those in Figure 3a. The orientation of longitudinal dunes and cross-stratification in the dunes indicated winds from the northwest and west-northwest. The source of the sand in the dunes was inferred to be fluvial sediment, partly glacial (outwash) and partly postglacial in age, in the river valley. Borns and Hagar (1965: 1243) thought the dunes likely formed "in a periglacial environment" but had no dates to support this. McKeon suggests a Bolling age (14– 14.4 ka; Figure 4).

In the Penobscot valley, most of the dunes are also east of the river, and thus downwind from the early-Holocene fluvial sediment in the valley. However, the presence of dunes west of the river (Figure 6c) suggests either other sources or other wind directions. Exposures were not good enough to permit any analysis of wind direction. In addition, growth of the Penobscot paleodelta, discussed next, suggests that the river was a sink for sediment, rather than a source at this time.

Penobscot paleodelta

For some period of time prior to ~ 9 ka, the Penobscot River was depositing sand in a paleodelta near the head of Penobscot Bay (Figure 1). The delta surface is presently ~30 m below sea level, well above the −60 m lowstand level reached at 12.2 ka. The base of the delta is at an elevation of about −45 m (Belknap et al., 2005, Figures 4–7). Relative sea level had risen to this level by ~12 ka (Figure 2), but then entered a period of much slower rise lasting from ~11.8 ka until ~8 ka (Kelley et al., 2013). The delta formed, in part, in response to this relatively slow rate of relative sea-level rise (Belknap et al., 2002, 2005). Two *Mya arenaria* shallow marine shells from channel sands within the delta yielded ages of 9.2 ± 0.1 ka (Barnhardt et al., 1997). Another growth-position *Mya arenaria* directly on top of the delta yielded a date of 8.5 ± 0.2 ka (Kelley et al., 2011). Belknap et al. (2005) and Kelley et al. (2011) attribute the abandonment of the delta to diversion of the Moosehead Lake drainage to the Kennebec River, thus reducing both discharge and sediment load. An increasing rate of sea-level rise after 8 ka would also weigh against deltaic deposition in the estuary (Belknap et al. (2005).

Terraces between Howland and Sunkhaze Rapids

Correlating terraces

We have mapped a number of terraces along the Penobscot River between Howland and Sunkhaze Rapids (Figures 6a–d and 7). Interpretation of these features is complicated by the presence of riffles that were not exposed when the terraces were formed.

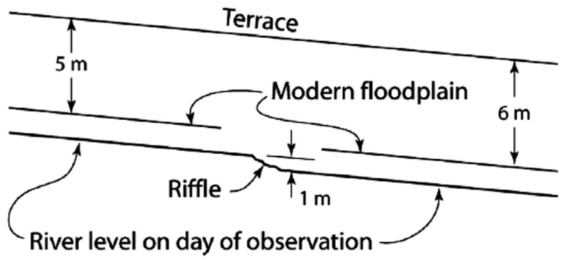

Figure 8. Sketch to illustrate procedure for taking drops over riffles and rapids into consideration in correlating terraces.

To take the late development of these riffles into consideration, we measured or estimated the present drops across them, and the present-day elevations of the terraces above the modern floodplain. Where a drop across a riffle is 1 m, the modern floodplain some distance up-river from the riffle will be 1 m higher than below it (Figure 8). Thus, for such a riffle, a terrace that is, say, 6 m above the level of the modern floodplain below the riffle will be only 5 m above the modern floodplain (maf) above it.

With the use of a numerical model (Brotchie and Silvester, 1969; Hooke and Ridge, in press), we also estimated that, owing to passage of the forebulge, the slope of the river at ~6 ka, when the highest terrace was formed (see below), was ~0.02 m/km (10%) steeper than it is today. This was also taken into consideration.

Greenbush high terraces and paleochannels

We have studied the terraces near the town of Greenbush (Figure 6c) most carefully, and thus begin our description of these features there.

The highest surface in this area is ~8.9 ± 0.4 maf. It lies between the river and some dunes to the east (Figure 6c) and is underlain by >4.5 m of well-sorted, sub-horizontally laminated fine sand with a few discontinuous silt-clay lenses, 2–10 mm thick. Bedding in the sand dips gently toward the river, suggesting that the dunes are the primary source of the sand. This is consistent with the fact that the surface cannot be traced either up- or downstream. We suspect this feature is a fan that once was graded to the next lower surface, the Sugar terrace described next.

A terrace on Sugar Island, herein dubbed the Sugar terrace, is 6.0 ± 0.3 maf. In one exposure, it is underlain by ~3 m of stratified sand, ranging in size from very fine to coarse, with one 0.2-m bed of clast-supported pebble gravel. This sediment, interpreted to be fluvial, overlies light gray clayey silt, interpreted as glaciomarine (Presumpscot Formation).

A terrace at 6.7 ± 0.2 maf is located immediately south of the above-described 8.9 m surface. This terrace is underlain by >4.3 m of moderately well-sorted, locally laminated, fine to medium sand with some lamellae composed of silt and clay. The section, again interpreted to be fluvial, also rests on marine silt and clay of the Presumpscot Formation. Making an adjustment for a riffle with a surveyed drop of slightly over 0.6 m, this terrace appears to correlate with the Sugar terrace (6.0 ± 0.3 maf + 0.6^{+} m ≈ 6.7 ± 0.4 maf.). A remnant of possibly the same terrace at ~6.4 ± 0.3 maf is located on the east side of the river, due south of Sugar Island. Here, the projected elevation of the Sugar terrace is ~6.0 ± 0.3 m. Still another remnant, varying from ~5 to 7 maf occurs on Olamon Island where the projected elevation of the Sugar terrace is ~6.1 ± 0.3 maf. We tentatively also correlate these latter two remnants with the Sugar terrace (Figures 6c and 7).

OSL ages of 6.7 ± 0.7 and 8.1 ± 0.9 ka were obtained from samples from depths of 1.2 and 3.6 m (7.7 and 5.3 maf, respectively) in the 8.9-maf surface (Figure 6c). A sample from a depth of 2.8 m (4.2 maf) in the terrace south of Sugar Island yielded an OSL age of 7.9 ± 1.1 ka. One from a depth of 2.5 m (3.2 maf) in the terrace on Olamon Island yielded an OSL date of 7.0 ± 0.6 ka. These dates and the one just above the 11.4-ka gravel in the Passadumkeag area (9.0 ka; Figure 6b) are plotted in Figure 9 together with one other in the Passadumkeag area that is discussed below. The systematic decrease in age with increasing height above the modern floodplain is interpreted as indicating that the material accumulated more or less continuously starting at or slightly before 9.0 ka. The rate of accumulation was ~2.4 ± 1.5 m/ka. Dividing the depth of an OSL sample by this sedimentation rate and subtracting this from the OSL age yields an estimate of the time of abandonment of the terrace surface as a site of active deposition. This calculation suggests that the 8.9 maf surface and the Sugar terrace were abandoned ~$6.4^{+0.4}_{-1.5}$ ka.

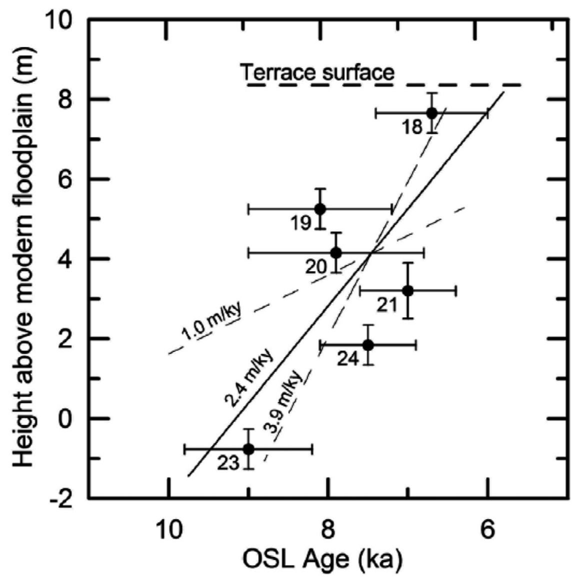

Figure 9. Plot of OSL age of sand samples from terraces against height of sample above modern floodplain suggests a sedimentation rate of 2.4 m/ka. Dashed lines are an estimate of the uncertainty in the slope based on standard rules of error propagation. Sample numbers from Table 1 are shown.

At many of these sites, the terraces are adjacent to dunes. The OSL dates from the terraces are, for the most part, younger than those from the dunes (Table 1). A ground-penetrating radar profile extending from the location of the 9.9/8.7-ka aged dune to the 6.7/8.1-ka terrace (Figure 6c) verified that the fluvial sediments lap onto the dunes. The grain size distributions of the two are quite similar (Figure 3), however, suggesting that the terrace sands were derived from the dunes.

Passadumkeag high terraces

At the Passadumkeag site (Figure 6b), a high area in the terrace southeast of Beaver Brook is ~6.1 maf. Northwest of Beaver Brook, the (presumably) same terrace is 5.5 ± 0.4 maf. Our projection suggests that the Sugar terrace should be at ~5.9 ± 0.4 m at Beaver Brook. Thus, we correlate these surfaces with the Sugar terrace (Figure 7).

The 6.1-maf surface is underlain by 1.2 m of medium to coarse sand grading upward to unstratified fine sand, interpreted to be fluvial in origin. This sand rests on marine clay of the Presumpscot Formation. East of the 6.1-m site, the surface material varies from coarse sand to tan silt, inferred to be fluvial and glaciomarine, respectively. Northwest of the 6.1-m site but southeast of Beaver brook, the sand gives way, without significant change in elevation, to very poorly sorted sandy silt with small pebbles, interpreted to be glacial till. Similarly, the materials underlying the surface northwest of Beaver Brook seem to vary from braidplain sands to till to marine clay. At present, we suspect that this relatively flat topographic surface is a thinly veneered strath terrace, on parts of which fluvial sands and gravels were either never deposited or from which they have been striped.

Howland high terraces

South of the confluence of the Piscataquis and Penobscot rivers, there are two larger and two small terrace remnants at ~7.4 maf (Figure 6a). The former are separated by a paleochannel at ~4 maf. On the western of the larger remnants, there is a small area of dune-like topography underlain by relatively uniform medium sand. Elsewhere, these surfaces are underlain by the material ranging from very fine sand to rare shows of small-pebble gravel. Locally, cobbles and boulders are present in gullies and in risers along the edges of these remnants, suggesting that the sand is relatively thin and is underlain by till or the early postglacial (>9 ka) gravel.

Another high terrace lies along the west bank of the Penobscot north of the confluence. This terrace, ~8.8 ± 0.3 maf, has been modified, perhaps extensively, for housing. It is underlain by poorly sorted very fine sand overlying gravelly silty sand. The latter has clasts up to small boulder size; it is probably the early postglacial gravel with additions of coarse lag from erosion of till. A high area near the terrace's southern end is underlain by fine sand with a size distribution very similar to those obtained from dunes (Figure 3a); this is interpreted to be a modified eolian dune.

Still another terrace, at ~8 maf and underlain by silty fine sand resting on till, lies along the east bank of the Penobscot south of the confluence (Figure 6a and b). This terrace extends for ~2 km along the river, dropping gradually to ~7 maf and then, over the next 0.3 km to ~3 maf. This last drop is presumably because of erosion. In the gap between this end of the terrace and the remnants of the Sugar(?) terrace NW of Beaver Brook (Figure 6b), there is a relatively flat surface at ~6.6 maf, underlain by till and marine clay. This we infer to be an extension of the strath(?) surface NW of Beaver Brook.

The distribution of these terrace remnants suggests that the confluence of the Penobscot and Piscataquis rivers was braided at the time when they were formed. The observation that some consist of a veneer of silty fine sand overlying till implies that they are straths.

Owing to the topography on which the Penobscot was superimposed as it extended itself southward, the river gradient between Howland and Passadumkeag at 7 ka was steeper than it was further south. Taking this into consideration, the expected height of the Sugar terrace at Howland is ~8.4 ± 1.2 maf. Thus, correlating the thin veneer of fine sediment with that beneath the Sugar terrace (Figure 7) is consistent with their being the highest surfaces in the area and, in at least two cases, appearing to be associated with eolian dunes.

Lower terraces

There are several lower terraces in Howland and between Howland and Sunkhaze Rapids (Figure 7). Because our OSL dates suggest that there was one period of aggradation, lasting from ~9.5 to ~6.4 ka, we infer that the lower terraces are all strath surfaces carved out of this valley fill. We describe them briefly next.

Extending south from the Sugar terrace yielding the 7.9-ka date in the Greenbush area is a surface at 1.5 ± 0.3 maf (Figure 6c and d). Other remnants of a surface at roughly this height occur on two islands south of Sugar Island (Figure 6c and d) and on the west bank of the river further south (Figure 6d). An old backhoe trench in the latter exposes alternating beds of granule gravel, very coarse sand, finer sand, and silt.

In the Passadumkeag area (Figure 6b), on the west bank of the Penobscot, and also on the west bank of Beaver Brook, there are terraces with surfaces at ~4 maf. These are underlain by locally weakly stratified fine to medium sand. A lower surface occurs at ~2.0 maf. An OSL date of 7.5 ± 0.6 ka was obtained from sand collected 0.3 m below the top of this latter surface (#24 in Figure 7).

In the Howland area (Figure 6a), there are surfaces at ~6 ± 1, 3–4, 1.5–3, and 1 maf, presumably reflecting pauses in downcutting from the Sugar terrace level. The most extensive of these is the ~6-m surface along the southwest side of the Piscataquis River and extending south from the confluence with the Penobscot. This surface is underlain by silt and very fine sand which, in turn, rests on till. South of the confluence it appears to be 1–2 m lower than the higher (7.4 maf) remnants to the east, but its elevation has not been accurately determined; here it slopes gently eastward, toward a paleochannel.

Paleochannels

Several linear wetlands, interpreted to be paleochannels, lie parallel to and east of the river in and south of Greenbush (Figure 6c and d). Cores were taken along transects crossing two of these. They revealed 1–2. m of peat overlying fine to coarse sand. Basal peat in the thalweg of the western one yielded a ^{14}C date of 9.5 ka (cal) and sand from just beneath the peat gave an OSL age of 7.5 ± 2.4 ka (Figure 6d). The latter is likely a minimum age because dose rate information from the overlying peat was not available; thus, the dates are reasonably consistent. Charcoal in the basal peat of the eastern paleochannel yielded a ^{14}C age of 3.3 ka. These are interpreted to be close minimum estimates of the time of abandonment of the respective channels.

Along the east side of the western paleochannel, there is a surface, underlain by poorly sorted fine to very coarse sand, that is probably a floodplain of the paleochannel. Along a level line crossing this paleochannel (Figure 6d: between the locations of the 7.9 and 7.7 ka dates), this surface is at ~2.2 maf, ~1 m above the adjacent wetland but 4 m below the younger Sugar terrace. This raises the question of why this channel apparently was not active during deposition of the sediment underlying Sugar terrace.

At the site of the 9.5-ka date, the peat is 1 m thick in the thalweg, so the bankfull depth was ~2 m, about half that of the present Penobscot River (estimated to be ~4 m, based on the difference between the gage height of the minimum observed discharge since 1902 and the gage height of the mean annual flood). The bottom of this channel is ~5 m above the present bed of the Penobscot. The relatively shallow depth suggests braiding. The peat in the eastern paleochannel is up to 2.5 m thick, but the elevation of the bog surface there is not accurately known.

In the Passadumkeag area (Figure 6b), at roughly the level of the mean annual flood, there is a linear wetland bordered by a surface at ~1.5 maf. On a coring transect, the peat was found to be ~1.1 m thick and was underlain by sand. This, too, is interpreted to be a paleochannel (with a bankfull depth of ~2.5 m), the bottom of which is ~4 m higher than the present bed of the Penobscot. Wood from basal peat in this paleochannel was dated to 4.1 ka.

In the Howland area (Figure 6a), there are two linear wetlands at elevations of ~4 and ~0 maf, interpreted to be paleochannels. The higher (eastern) one is bordered by a surface at ~5 maf and the lower one by a surface at ~1 maf. During an extreme flood, the Piscataquis River could, at present, overflow into the latter.

Interpretation of features between Howland and Sunkhaze Rapids

An interpretation that honors all the presently available data and observations along the river between Howland and Sunkhaze Rapids is as follows:

Prior to ~12 ka, the river was flowing over glacial materials and its gradient was determined, largely, by the topography left by the ice sheet. It was not graded. The gravels beneath the sands of the present floodplain suggest that initially, while discharges were still high, the river cut down into the glacial deposits. During this phase, it may have left the straths cut on glaciomarine sediment and till beneath some of the higher terraces.

Between Howland and Passadumkeag, the slope of the land, and hence of the river, was steeper than between Passadumkeag and Sunkhaze Rapids (Figure 7). Below Sunkhaze Rapids, bedrock inhibited downcutting that might otherwise have occurred as relative sea level dropped to its postglacial −60 m low.

Between ~12 and ~9.5 ka discharge in the river was reduced, first by retreat of the ice from northern Maine, then by the drying climate, and finally by diversion of flow from the Penobscot to the Kennebec drainage as Moosehead Lake tilted southward. The drying climate led to less vegetation and hence to extensive eolian activity, with sand likely blowing in from the west. The load of the river shifted from gravel to fine sand, and with the diminished discharge the river became overloaded. Between Passadumkeag and Sunkhaze Rapids, it began to aggrade, but the steeper slope between Howland and Passadumkeag limited aggradation in that reach. The aggradation led to activation (or reactivation) of several channels, now preserved as peat-filled paleochannels with sandy beds. Significant amounts of the sand were also transported to the sea where the rough balance between sea level rise and rebound led to a relatively constant sea level and aggradation of the Penobscot paleodelta.

At ~10 ka, the western paleochannel was carrying coarse sand and granules. Shortly thereafter, however, aggradation near Greenbush may have blocked this channel. The eastern one, however, continued to carry significant flow. At ~6 ka, in response to increasing precipitation, stabilization of dunes by vegetation, and perhaps some continued differential rebound, the river began to downcut, leaving its braidplain as the Sugar terrace. The observation that the Passadumkeag and eastern paleochannels remained active for a time suggests that the sediment, although relatively fine, was still carried largely as bedload. When one anabranch of a bed-load-dominated braided river cuts down slightly, others become sediment starved and are able to downcut, thus keeping up with the first. Eventually, however, first the Passadumkeag (4.1 ka) and then the eastern (3.3 ka) one were abandoned. Local pauses in the downcutting, perhaps in part a result of riffles over bedrock or over lags of coarse cobbles, led to the formation of local terraces.

While this model accommodates known features and also assumes our dates are all valid, it undoubtedly has flaws. It does, however, provide a starting point for future study. Particularly puzzling are the old dates from the eastern paleochannel.

The river downstream of Sunkhaze Rapids

The gradient of the Penobscot from Sunkhaze Rapids to tidewater at Bangor is nearly five times the gradient between Passadumkeag and Sunkhaze Rapids (Figure 7). Thus, less aggradation has occurred in this reach since the transition to fine-grained sedimentation at ~10 ka. Furthermore, sedimentation at the Mackowski and Blackman Stream archeological sites, both of which are in the present floodplain, appears to have been only intermittently continuous, with periodic intervals, lasting up to a millennium, during which accumulation was slow enough to allow soil formation (Kelley and Sanger, 2003).

Summary and conclusion

As the Laurentide ice sheet retreated across the Penobscot Lowland and up the East Branch, the area was isostatically depressed and inundated by the sea. Retreat across the Lowland was rapid, averaging ~150 m/a. Numerous small moraines, likely annual, were formed within ~100 km of the coast (Eusden, 2014). These are unmistakable on LIDAR, but were almost universally overlooked during earlier fieldwork. Further inland, large esker heads mark longer pauses in the retreat.

The Penobscot River was reborn ~14 ka near the confluence of the Seboeis River with the East Branch. Between ~14 and ~12 ka, the Penobscot was extended, both to the north as the ice retreated and to the south as isostatic uplift exceeded sea level rise. Discharges were larger than at present as ice was still melting in the headwaters and the diversion of flow from the Penobscot basin into the Kennebec River at Moosehead Lake had not yet occurred. These high discharges likely first eroded some of the glacial cover and then transported and deposited gravel that was eroded from the Katahdin esker and from a series of deltas along the East Branch (Hooke and Hanson, in preparation; Stone, 1899: 15).

In the Penobscot Lowland, the river's gradient and course were determined by the topography left by the ice sheet. Initially, the river cut down through glacial materials, leaving coarse gravel lags on strath surfaces between Howland and Passadumkeag and in the valley bottom further downstream. Sometime between 12 and 10 ka, probably during the early phases of the Younger Dryas, eolian sand began to accumulate. Between Howland and Sunkhaze Rapids, the sand formed dunes on either side of the river, and in lower-gradient stretches between Passadumkeag and Sunkhaze Rapids, it resulted in aggradation of the river, burying the outwash gravel. Sand also began to accumulate near Howland and downstream of the Sunkhaze, but in these locations steeper river gradients limited aggradation. Sand also accumulated in a paleodelta in Penobscot Bay. The cold dry climate of the Younger Dryas, Preboreal, and early-Holocene, the final melting of ice in Maine, and the diversion of Moosehead Lake overflow into the Kennebec River all contributed to this change in regime. Eolian activity in the mid-section of the Penobscot lowland continued to at least 8 ka and river aggradation between Passadumkeag and Sunkhaze Rapids until ~6 ka. The latter formed a braidplain, now the Sugar terrace, ~6 m above the present floodplain. During the aggradation, several paleochannels were formed (or reoccupied). Between Howland and Passadumkeag, the higher gradient of the river limited aggradation.

As the climate became wetter and warmer in the late- Holocene, after ~6 ka, the dunes became stabilized by vegetation, the sediment load likely decreased, the river gradient may have increased slightly because of continued differential rebound, and downcutting began. Anabranches of the river were progressively abandoned. Pauses in the downcutting are reflected in strath terraces cut into the Sugar terrace. Upon eroding through the early-Holocene sand deposits, the river reencountered the old outwash gravels, the marine clay of the Presumpscot Formation, glacial till, and locally bedrock. Its profile is now stepped, with gentle, generally gravel-bedded reaches between bedrock ribs that form rapids. Downstream from the Sunkhaze, aggradation appears to have slowed, but to have continued intermittently to the present. Sediment transport is presently low, as indicated by the lack of significant accumulation behind dams (Dudley and Giffen, 2001).

The complexity of the interplay among sediment load, discharge, and initial topography in controlling river morphology and history is well illustrated by the Penobscot River and its terraces. Sediment load depends on the nature of the materials present, and both sediment load and discharge depend on climate. In young river systems like the Penobscot, both load and discharge are imposed on an initial topography, the effects of which on the morphology are substantial and enduring.

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