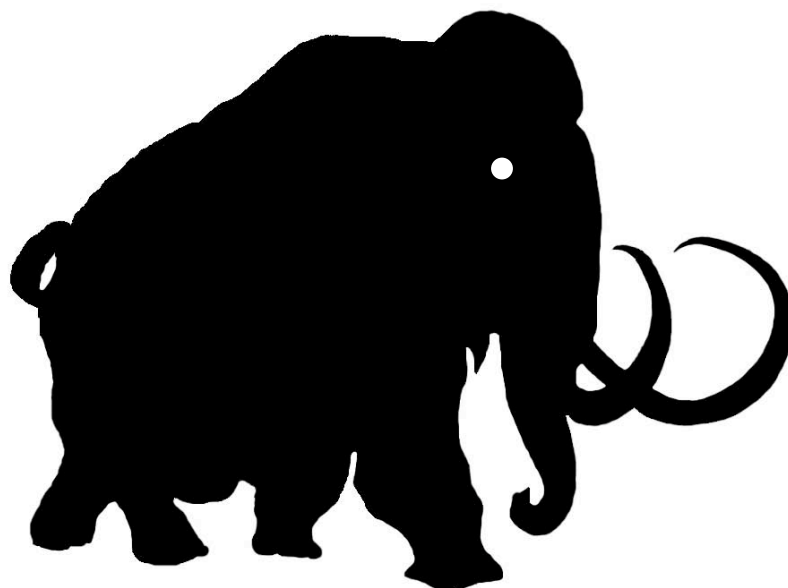


# Glacial and Archeological Features of the Penobscot Lowland, Central Maine

A guidebook prepared for the 69<sup>th</sup> annual field conference, of the Northeastern Friends of the Pleistocene, June 2-4, 2006, Orono, Maine



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Orono, ME 04469



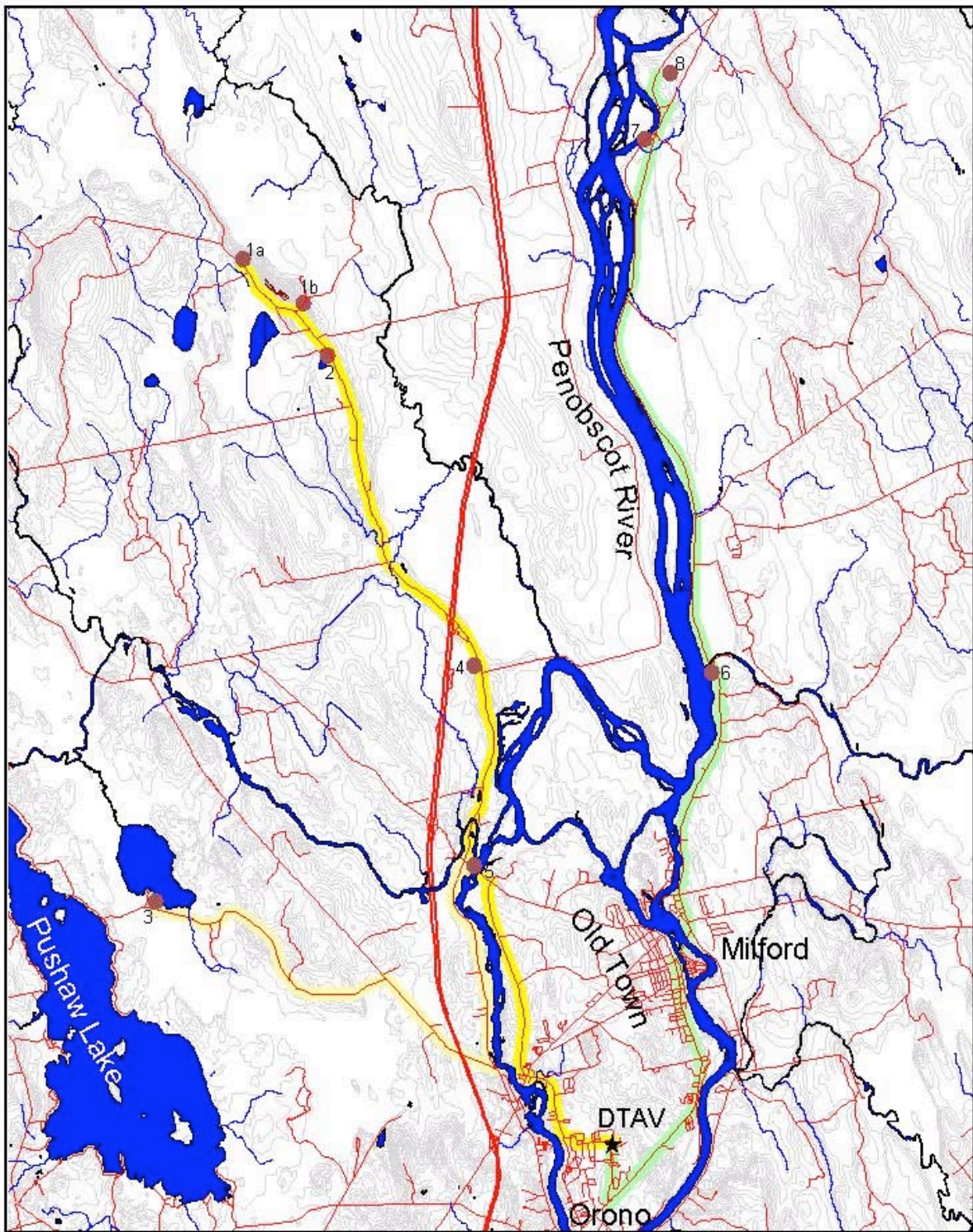
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## **Acknowledgments**

I am indebted to many people for making this guidebook a reality. My co-authors, of course, were indispensable in getting it together. I hope they will forgive me my heavy-handed editing; I have tried to preserve their intent while integrating their contributions into the whole. Deserving of special mention is Alice Kelley, on whom I have leaned extensively for help and input. Others who have contributed time and expertise are Ann Diefenbacher-Kahl, Dianne Perro, and our student assistants, Peter Leach, Chris Miller, and Bert Pelletier, who helped collate the guidebook and will be indispensable on the trip itself. I am also indebted to Mark Carroll, Aubrey Engstrom, Parker "Chip" Laite, William and Francine Mackowski, Gayland Madden, and David Wright and the town of Old Town for permission to visit sites on their properties.

*R.LeB.H*



FOP field trip route - 2006

1 0 1 2 3 Miles

- i -



## Field trip time schedule

### ***Saturday, June 3***

7:45 Assemble at Edith Patch turn out  
8:00 Leave Edith Patch  
8:30 Stop 1a: Esker stop (*RLH*)  
9:30 Leave Stop 1a  
9:45 Stop 1b  
11:00 Leave Stop 1b  
11:15 Stop 2: Mansell pond (*George Jacobson*)  
12:00 Leave Mansell Pond  
12:30 Stop 3: Lunch and Pushaw Bog Complex (*Alice Kelley and Dave Sanger*)  
2:00 Leave Stop 3  
2:30 Stop 4: Presumpscot (*Hal Borns*)  
3:15 Leave Stop 4  
3:30 Stop 6: Gilman Falls (*Alice Kelley and Dave Sanger*)  
4:30 Leave Gilman Falls  
5:00 Arrive in Orono

### ***Sunday, June 4***

7:45 Assemble at Edith Patch turn out  
8:00 Leave Edith Patch  
8:30 Stop 7: MacKowski Farm site (*Brian Robinson*)  
9:30 Leave Stop 7  
10:00 Stop 8. Braidplain Stratigraphy (*Roger Hooke*)  
10:45 Leave Stop 8  
11:00 Stop 9  
11:45 Leave Stop 9  
12:00 Lunch stop  
1:00 Leave lunch stop  
1:30 Arrive in Orono (field trip ends)

# Glacial and Archeological Features of the Penobscot Lowland, Central Maine

## A guidebook prepared for the 69<sup>th</sup> Annual Field Conference, of the Northeastern Friends of the Pleistocene, June 2-4, 2006 Orono, Maine

### INTRODUCTION

#### Late glacial history:

The Laurentide Ice Sheet began its retreat from the edge of the continental shelf off Maine somewhat before 22,100 (cal) years ago. By about 16.1 (cal) ka the ice margin in the Penobscot Bay region stood just inland from the present coastline (Fig. 0.1) (Borns *et al.*, 2004). For the next ~200 years it fluctuated back and forth, retreating slowly. It then began to retreat rapidly, and by 14.0 (cal) ka the Penobscot lowland was ice free and the margin was near Medway.

When the ice margin stood near the coast, the land was depressed and the ice terminated in the sea. The marine limit along this part of the coast is ~70 m above present sea level (m asl). Deposition along the ice margin occurred in the form of grounding-line moraines, submarine fans, and massive deltas. As the ice retreated from the coast, the sea maintained contact with it in lower areas, inundating the lowland (Fig. 0.1). The distribution of fossiliferous marine silts and clays of the Presumpscot formation is the primary line of evidence for this inundation. Some investigators have proposed that a marine embayment (a calving bay) formed in the Penobscot lowland, leaving ice on higher ground on either side (Shreve, 1985a; Dorion *et al.*, 2001).

The marine-embayment scenario, while supported by radiocarbon dates, has been questioned (*e.g.* Smith, 1985). Among other things, it is not consistent with the courses of eskers. The Katahdin esker (Fig. 0.1), for example, begins north of Medway and follows a southward course for ~80 km. It then veers abruptly to the southeast, leaves the Penobscot lowland, and trends across higher ground before

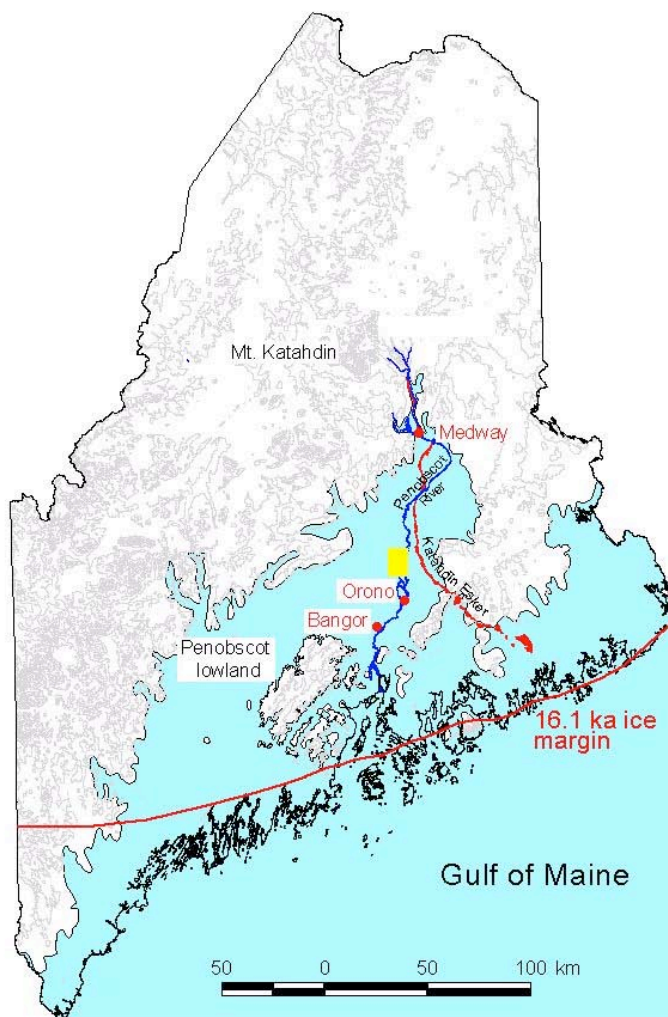


Figure 0.1. Map of Maine showing marine inundation in blue and location of field trip (Greenbush Quadrangle) in yellow. Gray shading is from contours at 150 m intervals and shows areas greater than 150 m above sea level.

ending near the location of a  $\sim 15.9$  (cal) ka margin. To divert the esker in this manner, the surface slope of the ice sheet on the eastern side of the lowland must have been to the southeast (Shreve, 1972, 1985a,b), requiring a higher ice-surface elevation in the lowland. Some small moraines along the eastern side of the Penobscot Bay trend northeast-southwest, which is consistent with the presence of an ice lobe in the lowland and not with a calving bay with residual ice on flanking higher ground. Furthermore, striations indicating ice flow toward the lowland from the adjacent higher ground are notably lacking (Smith, 1985; Dorion *et al*, 2001).

As the ice margin retreated up the East Branch of the Penobscot, the sea continued to flood in. The marine limit here is  $\sim 120$  m asl. Submarine fans and deltas formed where subglacial esker-building water emerged from beneath the ice sheet. Once the margin retreated above sea level, a proglacial outwash plain or braidplain formed, connecting the ice sheet with the sea for some time.

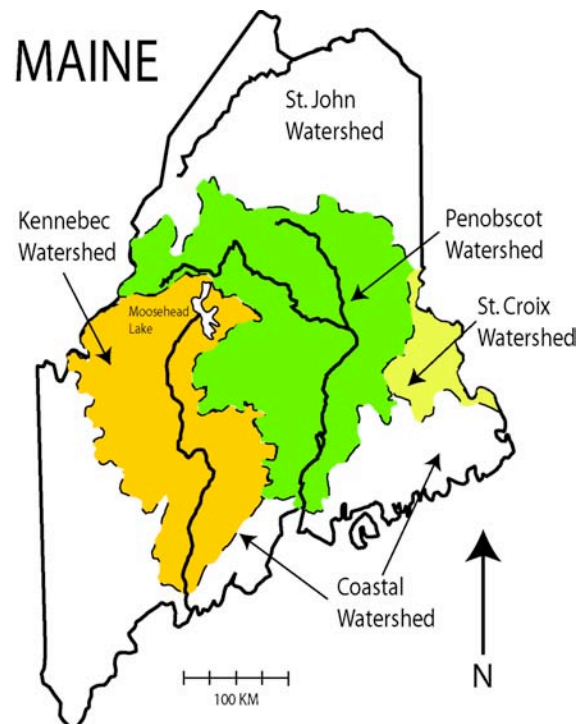
As sea level fell the braidplain was incised, leaving a series of alluvial terraces. During reconnaissance mapping, we've identified terraces at  $\sim 2$ , 4, 6, and 10 m above the present floodplain. The continuity of the terraces has not been established, nor have elevations been determined more precisely than can be done by hand leveling.

### The Modern Penobscot River

The Penobscot River watershed encompasses  $24,300 \text{ km}^2$  of central Maine (Fig. 0.2). It is the largest river system in the state, when defined on the basis of drainage area and average discharge ( $465 \text{ m}^3/\text{sec}$ ) (U.S. Army Corps of Engineers, 1990). The river is more than 160 km long, from the headwaters of either of the two major tributaries, the East and West Branches, to its mouth in Penobscot Bay. The Penobscot watershed is bounded to the east by that of the St. Croix, and to the west by Moosehead Lake and the Kennebec River watershed.

The present day Penobscot River is a result of its bedrock framework and surficial history. Kelley *et al.* (1988) divided the drainage into four regions on the basis of distinct geologic and geomorphic features (Fig. 0.3). The West and East Branches are designated as the **Headwaters Division**, an area of steep gradients in mountainous terrain. This division's most prominent feature is the Katahdin

Figure 0.2 Penobscot, Kennebec, St. John, and St. Croix drainages



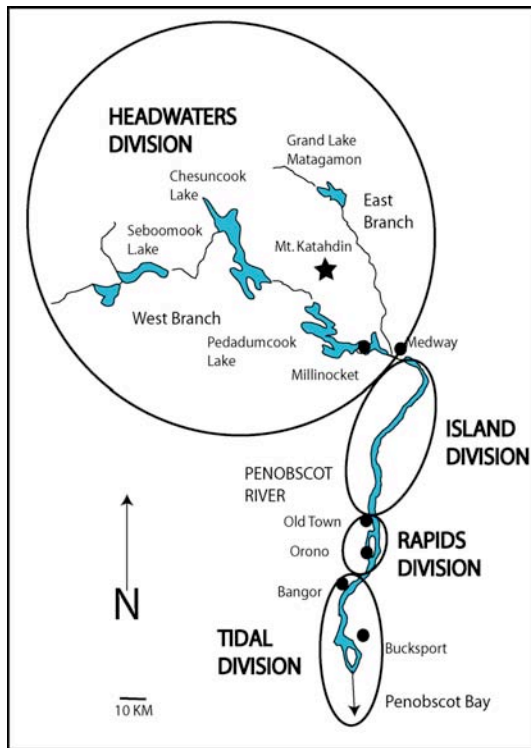


Figure 0.3 Geomorphic divisions of the Penobscot River

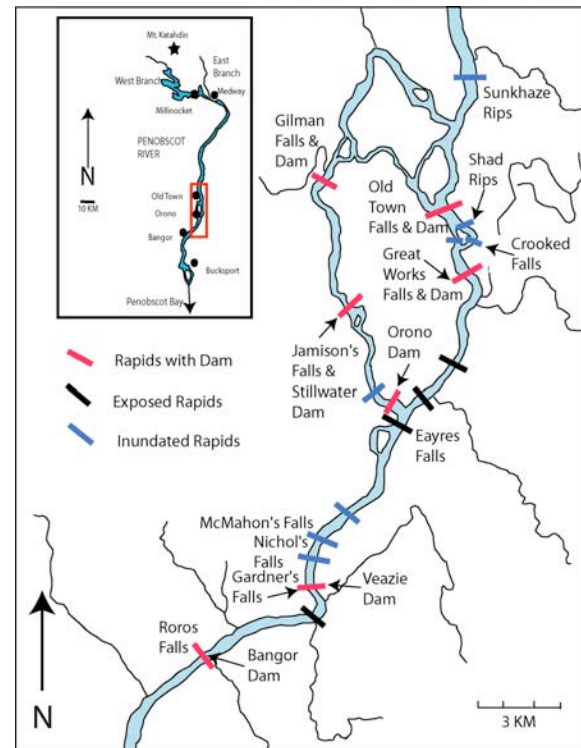


Figure 0.4 Lower Island Division and Rapids Division showing location of rapids. Modified from Treat (1820).

Massif, a Devonian granite pluton associated with felsic volcanics (Osberg, *et al.*, 1985). Other mountains in the region are formed of erosion-resistant contact metamorphic rocks, while lakes are associated with coarser-grained granite bodies (Hanson and Caldwell, 1989). Low- to medium grade metasedimentary rocks underlie the valleys (Hanson and Caldwell, 1989). Exposed rock surfaces are common at higher elevations, but topographically lower areas are mantled by till. Outwash deposits, including the terraces mentioned above, are found in valleys. A broad area of ribbed moraine exists to the south of the Katahdin massif (Thompson and Borns, 1985). Till in the Headwaters Division is derived primarily from granitic sources, and has a sandy matrix. The Katahdin esker heads in this region. This high-relief interior portion of the drainage provides snowmelt to the annual spring freshet of the Penobscot.

The **Island Division** is a low-gradient area that encompasses the region from Medway to Old Town, and is characterized by a series of islands and rapids flanked by a wide floodplain. Fine-grained, lower Paleozoic metamorphic rocks of the Vassalboro and Madrid Formations outcrop throughout most of this area (Osberg *et al.*, 1985). The easily-eroded strata of these formations result in a broad valley, with gently rolling topography. Resistant strata and boulder lags formed by erosion of moraines create rapids. Surficial deposits in this portion of the drainage include eskers, glacial outwash, till, terrace deposits, and modern alluvium in addition to the glaciomarine Presumpscot Formation which drapes much of the landscape below 60 m asl (Bloom, 1960, 1963; Thompson and Borns, 1985, Hooke and Borns, pers. comm., 2003). Modern swamps and bogs are

ubiquitous in the low relief portions of the drainage, while numerous lakes are present in a region of granite and gabbro intrusions to the east (Thompson and Borns, 1985; Osberg *et al.*, 1985).

The reach from Old Town to Bangor, the **Rapids Division**, has an increased gradient in comparison with the Island Division. It includes a number of bedrock rapids, created by erosion-resistant layers in the Silurian/Ordovician Vassalboro Formation. These transverse bedrock outcrops form a series of local base levels. Dams are built on these bedrock “sills” in several locations in both the Island and Rapids Divisions. These dams create large headponds, dramatically changing the river's appearance and drowning numerous rapids (Treat, 1820) (Fig. 0.4). As in the upstream Island section, the Presumpscot silts and clays mantle much of the landscape. Where the Presumpscot is present, topography is low and rolling, with tributary streams occupying gullied valleys. Till is exposed at the surface at higher elevations, and commonly has a thick, clay-rich matrix as a result of its derivation from fine-grained metamorphic rocks. Eskers are common. Well-developed flights of terraces are present in Old Town, Orono, Bangor, and Brewer, but do not seem to be related to the terraces in the upper part of the Island Division or the Headwaters Division. South of Orono, the river has formed bluffs up to 12 m high by downcutting through the Presumpscot Formation and till.

The section from Bangor to Penobscot Bay is tidally influenced, and is identified as the **Tidal Division**. South of Bangor, the modern head of tide, the lower Penobscot flows through a region of low mountains formed by granitic plutons, volcanic rocks, and more resistant high-grade metamorphic rocks. Bedrock cliffs confine the river in several locations. Bluffs, composed of till and of glaciomarine and glaciofluvial deposits, are commonly up to 12 miles high. Rapids and falls are absent, probably drowned by rising sea level, giving the river a more mature appearance in contrast with the youthful, deranged setting upstream. Fringing salt marshes are developed in the mouths of tributaries and small indentations in the shoreline. Immediately south of Bucksport, the river divides into the main channel and the Eastern Channel to flow around Verona Island. Although mantled by glacial deposits, Verona Island is bedrock-cored so the river tends to have rocky banks. Somewhat south of Verona Island the river widens dramatically, forming Penobscot Bay.

## Cultural History of the Central Penobscot River Valley

The cultural history of Maine has been divided into four major periods: Paleoindian, Archaic, Ceramic (Woodland), and Contact (Historic), with subdivisions present within each period (Table 0.1). The periods and their subdivisions are recognized on the basis of diagnostic artifacts and/or by radiocarbon chronology. Chronological divisions of the archaeological periods vary slightly from author to author. Divisions used herein are drawn from Sanger (2005). Ceramic period subdivisions are after Petersen and Sanger (1991). The recognition of each of these subdivisions is based on changes in the technology, materials, and styles employed in the creation of stone tools and ceramic vessels.

The Paleoindian period represents the first recognized occupation of New England and the Canadian Maritimes, and extends from 11,000 to 9,500 BP. The Early Paleoindian tradition is identified by the presence of the “fluted point”, an elongate biface having a distinctive longitudinal scar, or flute, created by the removal of a channel flake from one or both faces of the tool (Snow, 1980; Spiess *et al.*, 1998). These tools are characteristically produced from high quality cryptocrystalline quartz or fine-grained volcanic rocks. The diagnostic Late Paleoindian tradition

artifact is the parallel-flaked, lanceolate biface. These artifacts are produced from the same lithic types associated with fluted points.

**Table 0.1**

DIVISION	TIME radiocarbon years BP
<u>Paleoindian Period</u>	
Fluted Point tradition	11,000-10,000
Late Paleoindian tradition	10,000-9,500
<u>Archaic Period</u>	9,500-3,000
Early Archaic	9,500-7,500
Middle Archaic	7,500-6000
Late Archaic	6,000-3,000
Late Archaic: Laurentian tradition	6,000-4,500
Small Stem Point tradition	5,000-3,800
Late Archaic: Moorehead phase	4,500-3,700
Late Archaic: Susquehanna tradition	3,800-3,000
<u>Ceramic Period (Woodland)</u>	3,000-400
Early (CP 1)	3,000-2150
Middle (CP2, 3)	2150-1350
Late (CP 4,5,6)	1350-400
<u>Early and Late Contact Period (CP 7)</u>	450-200

## Cultural periods in Maine

No Fluted Point tradition artifacts have been recovered in the Penobscot drainage. Material from this time period has, however, been found in association with chert outcrops at Munsungan Lake, to the north (Bonnichson, 1981; 1982) and the Searsmont, Maine area to the south (Cox *et al.*, 1994). Sites of this time period are widely distributed in southern Maine and New England. Late Paleoindian parallel-flaked points have been recovered from sites within the central Penobscot Valley, as well as other locations in Maine.

The transition to the Archaic period is marked by an abrupt change in lithic technology in Northern New England and the Canadian Maritimes. Fluted and parallel-flaked bifaces disappear from the archaeological record. Nonlocal rock types used in the manufacture of Paleoindian tools appear less frequently, and are generally replaced by locally available sedimentary and metasedimentary rock types. Ground stone tools are first seen in the New England and Maritimes archaeological record, and range from large celts and adzes to thin and delicate slate “bayonets”. Regional styles appear to supplant the earlier broad uniformity of form and technology.

The Archaic period is also associated with the “Red Paint Indians”. Popularized by the excavations of Willoughby (1898, 1935) and Moorehead (1922) in the early 20<sup>th</sup> century, this term is used to describe burials characterized by inclusion of red ochre and a distinctive set of ground stone grave goods. These features often lack human remains due to the acidity of the soils in interior New England and the Canadian Maritimes. Burial ceremonialism using red ocher is recognized throughout the Archaic, but flourishes at the end of the period (Robinson, 2001). Several “Red Paint” cemeteries have been identified in the central Penobscot Valley.

Ceramics appear in the regional archeological record at approximately 3,000 BP, and persist from this point through the Historic period. The sudden and striking appearance of these artifacts marks a major innovation in regional technology, and gives the Ceramic period its name. Although the term, Ceramic period, is widely used in Maine and the Canadian Maritimes, “Woodland” or “Maritime Woodland” is also used to describe the same interval. Subdivisions (CP1-CP6) within the Ceramic period are based on changes in manufacturing styles, decoration, and surface treatment (Petersen and Sanger, 1991). Vessels are made from slabs or clay coils, and can be thick or thin-walled depending on time of production. Temper materials include sand, rock fragments, crushed shells, and fiber. Surface decoration ranges from cord or fabric impression to stamping, smoothing, or the presence of punctuations. Rim forms vary through time, being simple, castellated, or collared. For most of the Ceramic Period, pots were generally cylindrical with pointed bases. More globular forms appear late in the period.

Ceramic period lithic artifacts are marked by a return to non-local lithic sources. Bifaces also tend to be smaller, and better suited to bow and arrow technology (Mack *et al.*, 2004). Groundstone tools are reduced from their earlier prominence, and are limited to pecked and ground celts and whetstones (Sanger 1979). In other portions of New England, the Ceramic period marks the advent of agriculture. There is no evidence of pre-Contact agriculture in the Penobscot Valley in either the archaeological or ethno-historical record

The arrival of Europeans in Maine marked a radical change in the lifeways of the native inhabitants. At first, interactions were limited to trade, primarily furs. In the Penobscot Valley, this began in 1580 when John Walker acquired 300 pelts from native inhabitants (Bourque, 2001). European goods, such as copper and iron pots and utensils, appear in the archaeological record at this time. Interactions between natives and Europeans continued, first amicable, then hostile. Epidemics of European diseases decimated Native populations, allowing European settlers to move more readily into Native lands. Fewer in number and with their social systems disrupted, Native groups became a marginalized part of Maine and New England society.

The changes brought about by European timber harvesting, agriculture, and industry also reshaped the land. Dams were built across the Penobscot and many of its tributaries, first to provide power for sawmills, later to produce electricity. The headponds associated with these dams drowned rapids, changing the appearance of the river, and changing processes in the Penobscot system. Sediment released by logging and agriculture swept into the streams. Dam construction, logging practices, and pollution changed patterns of fish migration. The river we see today is much different from that used by the land’s original inhabitants. This changed landscape and ecological system cannot be used as an analogue for investigating the region’s archaeological record.

## STOP 1: ALTON ESKER (*ROGER HOOKE*)

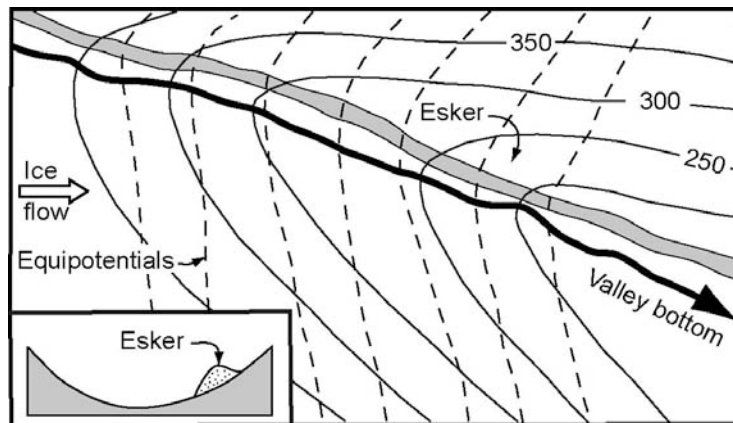
### Esker Physics

A classic paper by Shreve (1972) provides a basis for understanding the paths that eskers take. Shreve showed that water deep in a glacier will move in a direction normal to planes of equal hydraulic potential. In the ice, these planes dip upglacier at an angle that is  $\sim 11$  times the slope of the glacier surface.

When water gets to the bed, it flows normal to contours formed by the intersections of the englacial equipotential planes with the subglacial topography. (This is precisely analogous to saying that water on a subaerial land surface flows downhill normal to topographic contours. Topographic contours in a subaerial setting are contours of equal potential.) One can visualize the pattern made by equipotential contours on a glacier bed by thinking of the dipping equipotential planes in the ice as contacts between rock strata, and imagining the outcrop pattern of these contacts on the topography.

An application of Shreve's model is shown in Figure 1.1. The figure depicts a valley (solid contours) trending south of east, at  $\sim 20^\circ$  to the easterly direction of the ice surface slope. Under subaerial conditions, a stream would flow down the axis of the valley. However, with the ice sheet in place, the contours of equal potential define a valley with its axis displaced up onto the side of the topographic valley. An esker formed beneath this glacier would thus run along the valley side, as shown in the inset.

Figure 1.1. Topographic map of a hypothetical valley trending diagonally across the direction of glacier flow. An esker formed in such a situation would be up on the side of the valley. (Modified from Hooke, 2005, Figure 8.25)



In an englacial or subglacial water conduit, conduit walls are melted by heat produced by viscous dissipation in the flowing water. This melt is balanced by closure of the conduit, driven by a slight difference between the pressure in the water and that in the ice (Röthlisberger, 1972; Shreve, 1972). The latter is approximately equal to the ice overburden pressure. Where basal ice is laden with debris, the melting releases sediment which then *may* form an esker.

### Segmented Eskers

Many eskers are believed to have formed in segments (e.g. Donner 1965; Banerjee and MacDonald, 1975; Hebrand and Åmark, 1989; Ashley *et al.*, 1991, p. 121; Bolduc, 1992, p. 115). It is commonly inferred that the segments were built near the distal ends of subglacial drainage systems, as DeGeer (1897) described, and that successive segments formed as the ice margin retreated. Left unexplained, however, is why tunnel deposits were not formed in the drainage system farther upglacier, or why the segments link up as nicely as they do. Banerjee and

MacDonald (1975, p. 152), for example, conclude that “subglacial tunnels extend themselves headward as the ice front retreats” but note that “the details of this extension are unknown.”

In many eskers in Maine, a typical segment begins as a relatively small ridge. The ridge increases in size downflow, at first slowly and then more rapidly, resulting in a tadpole shape (Fig. 1.2). The (tadpole) head reflects a period of time during which the margin position was relatively stable.

Near the margin, ice thins and the glacier surface slope becomes steeper. The former decreases the pressure in the basal ice. The latter increases the potential gradient driving the flow, thus increasing the energy dissipation in the flow and hence melt rates on conduit walls. These two tendencies, combined, are inferred to have led to melt rates which exceeded closure rates by an increasing amount as the ice thinned toward the margin. The consequent increase in size of the conduit would have resulted in a decrease in water velocity, and hence in deposition of coarser sediment in the tunnel portal, forming the head. As this part of the Laurentide margin terminated in the sea, finer material that was flushed past the portal was deposited in submarine fans or, where

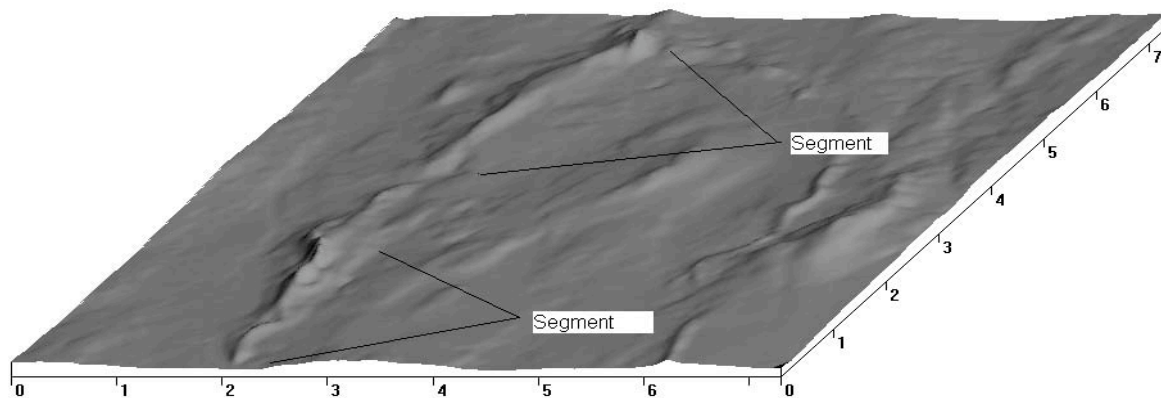


Figure 1.2. Oblique view, looking downglacier, of two segments on the Katahdin esker southeast of Olamon, Maine. Based on a 30 m DEM. Scale in kilometers.

the margin was stable long enough, in deltas. The finest material settled out still further offshore, contributing to the Presumpscot Formation. As the margin retreated, successive segments were formed. The head of each younger segment, together with its associated fan or delta, commonly laps onto and buries the tail of the previous one, like shingles. In the Katahdin esker, some 16 km east of the Alton esker, I have identified ~30 segments, averaging ~5 km in length. [This interpretation of the Katahdin esker differs substantially from that proposed by Shreve (1985a,b).]

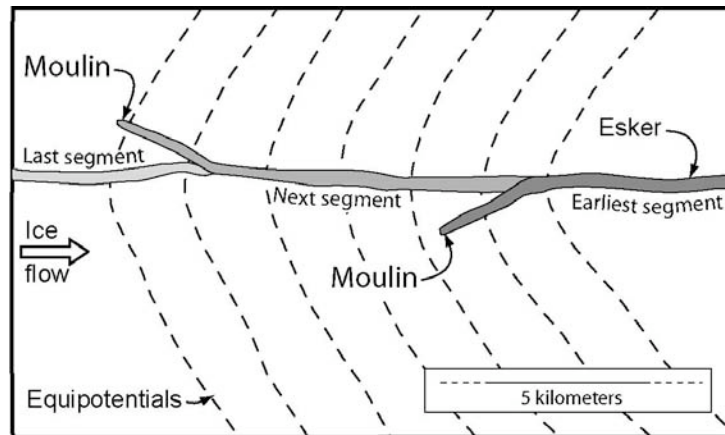
In a meticulous study of the lithologic composition of gravels in several eskers in Labrador, Bolduc (1992, p. 114f) found that, in some instances, vertical variations in lithology were reflected in longitudinal variations along the esker crest. Thus, the composition of basal beds in one place was similar to that of beds at the esker crest 5 to 10 km down flow. This led her to suggest the shingle analogy that I used above. Headward growth of the esker along some form of pre-existing subglacial drainage system is implicit in this analogy.

Shilts (1984, p. 218) thought that eskers in the Keewatin district, west of Hudson Bay, also grew headward. He proposed that the eskers are a reflection of “... an integrated system of drainage channels developed *on the surface* of the glacier...” (italics mine) and that “...meltwater plunge[d] to the base of the glacier to flow in a subglacial tunnel the last few kilometers of its course...” The new moulins, he thought, probably formed at intervals along the course of the

superglacial stream (W. W. Shilts, e-mail comm., 2/06). A problem with this model is that the surface drainage would not necessarily follow the course defined by the potential field at the glacier bed, and new moulins would not necessarily form in precisely the right place to result in a continuous esker. Thus, one would expect to see barbs representing situations in which the water arrived at the bed up on the side of a potential valley (Fig. 1.3). The next segment would then join the earlier one some distance down flow from the base of the moulin that initiated the earlier one.

In short, although it is generally agreed that many eskers were built headward in onlapping segments, heretofore we have not had a satisfactory explanation of how they maintained their continuity as they were extended headward.

Figure 1.3. Plan geometry of a hypothetical segmented esker built by water that reached the bed through moulins. It is assumed that successive moulins are formed as the ice sheet retreats, that the esker begins at the bottom of the moulin, and that it subsequently connects with the previously-formed segment. Moulins are unlikely to form exactly on the axis of the potential valley. Thus, barbs are expected.



## Alton Esker Stop

At this stop, we will look at the transition from tunnel to proglacial submarine fan deposition in an esker head. The transition takes place in three gravel pits, but owners of the central one would not give us permission to enter the pit. In the first pit (Sargent and Sargent), we will stand on the gravel core of the esker and look at a sand drape. The high point of the esker in the distance, presently 105 m above sea level, was close to, if not at the marine limit at ~14.5 (cal) ka.



Figure 1.4. Interbedded sands and gravels draped over an esker core in the central pit.

Figure 1.5 shows a longitudinal cross section of an esker ~65 km west of us, where interbedding of esker gravel and fan sand is well displayed. Figure 1.6 is a detail of the interbedding in the same part of the esker but taken a year earlier.

We will then proceed to the third (Engstrom) pit.



Figure 1.5. Interbedded sands and gravels at a transition from tunnel to fan deposition in a longitudinal section in an esker near Unity, ME.



Figure 1.6. Detail of exposure in Figure 1.5 but taken a year earlier

The esker gravel core that used to underlie the sand in this pit has already been carted away, leaving a trench in the fan. This trench extends some 300 m further south. After looking at the active pit, we will walk south along the former flank of the fan to view some glaciotectionic structures, presumably formed during an oscillation in the margin during which the glacier advanced, deforming the fan.

## Keeping Esker Segments Aligned

In a polar glacier, there is a temperature gradient at the bed that conducts heat upward into the ice. In places far from the ice margin, where the climate at the glacier surface is cold and little surface melting occurs, any water at the bed will have been derived from geothermal and frictional heating. These are both distributed sources, so the conduit system transporting this water should be distributed. That is, it will consist of many small broad low anastomosing channels. Numerical modeling of the sector of the Laurentide Ice Sheet in Maine (Hooke and Fastook, unpublished) suggests that, ~30 km from the margin, if such channels occupied 20% of the bed, they would be ~12 mm high and the water speed in them would be ~30 mm s<sup>-1</sup>. These appear to be reasonable values.

Now suppose that heat dissipated by the flowing water begins to melt the roof of an anabranch of this network, and that the flow thus becomes deeper. The resulting increase in discharge in the deeper anabranch will increase the energy dissipation there, and *under temperate ice* this will result in further melting. Owing to this positive feedback, Shreve (1985a, p. 642) suggested that the resulting conduit would become “sharply arched.” The conduit roof, however, must remain at the pressure melting point. Thus, *in cold ice* the temperature contours immediately above the conduit must be deflected upward relative to those a long distance from it (Fig. 1.7). Far above the conduit,

however, the contours will not be affected. Thus the temperature gradient above the conduit will be higher than elsewhere along the bed, and heat loss from the conduit will be greater. This will inhibit melting of the conduit roof, and should strongly dampen the tendency for a broad flat conduit to evolve into a sharply arched one by melting of the roof. Without such melting, sediment will not be released from the ice, and eskers will not form.

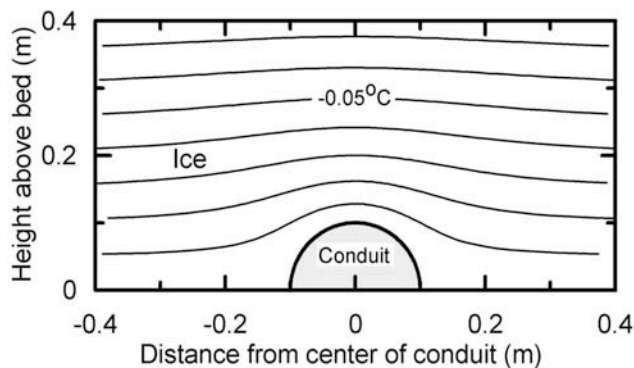


Figure 1.7. Calculated temperature distribution in cold ice above a semicircular conduit at the pressure melting point. Basal gradient, due to geothermal and frictional heating, is taken to be  $-0.20\text{ }^{\circ}\text{C m}^{-1}$ .

Toward the margin, the surface slope of the ice sheet increases, increasing energy dissipation in the water, and the temperature gradient upward into the ice decreases. Combined, these two factors increase the chance that a sharply arched conduit will develop. Thus, beneath a cold continental ice sheet, eskers may form only near the margin.

In short, water may well reach the bed of a cold continental ice sheet through moulins, as Shilts and many other glacial geologists have long assumed. Recent studies (Zwalley *et al*, 2002; Boon and Sharp, 2003) have documented that surface meltwater can reach the bed of a cold glacier by propagation of water-filled crevasses. Upon reaching the bed, however, the Pleistocene water may have encountered a distributed network of broad low conduits, such as would be characteristic of situations, some distance from the margin, where temperature gradients in basal ice are high enough to inhibit formation of arched conduits. The water may have spread out in this network and not coalesced to form a single sharply-arched conduit until it had flowed some distance, perhaps many kilometers, downglacier, to a point where potential gradients driving the flow were higher and basal temperature gradients lower. Surface meltwater could thus have been involved in the construction of continuous segmented eskers without leaving the barbs shown in Figure 1.3.

## STOP 2. Mansell Pond (George Jacobson)

Because moisture is such an important factor for biological activity, our understanding of long-term changes in vegetation and climate is greatly enhanced by information about paleohydrology.

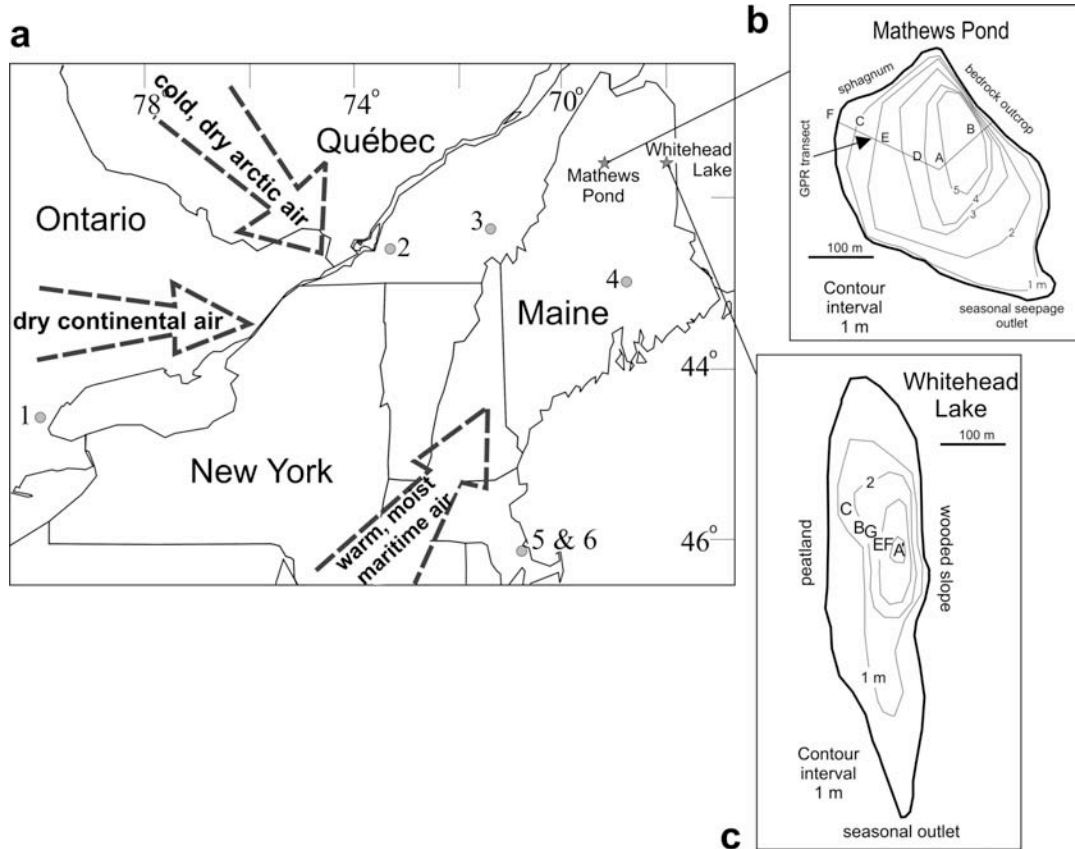


Figure. 2.1. Locations of Mansell Pond (#4 on map), Mathews Pond and Whitehead Lake. Air masses follow Bryson's work. From Dieffenbacher-Krall and Nurse (2005).

During the past several decades, researchers have applied and further developed the methods developed by Prof. Gunnar Digerfeldt (Department. of Quaternary Geology, Lund University, Sweden) in reconstructing the history of lake levels. Digerfeldt pioneered the approach of examining a transect of sediment cores from deep to shallow water in order to trace the position of the shoreline through time. Subsequent studies (e.g., Dieffenbacher-Krall and Halteman, 2000) have refined the methodology in useful ways.

Mansell Pond, in central Maine (Fig. 2.1) is one of several that have now been studied in this region to determine the temporal and spatial patterns of changing water balance during the Holocene (Almquist *et al.* 2001). Subsequently, research projects at Mathews Pond and Whitehead Lake in northern Maine (Fig. 2.1) have provided important additions to the regional understanding (Dieffenbacher-Krall and Nurse, 2005).

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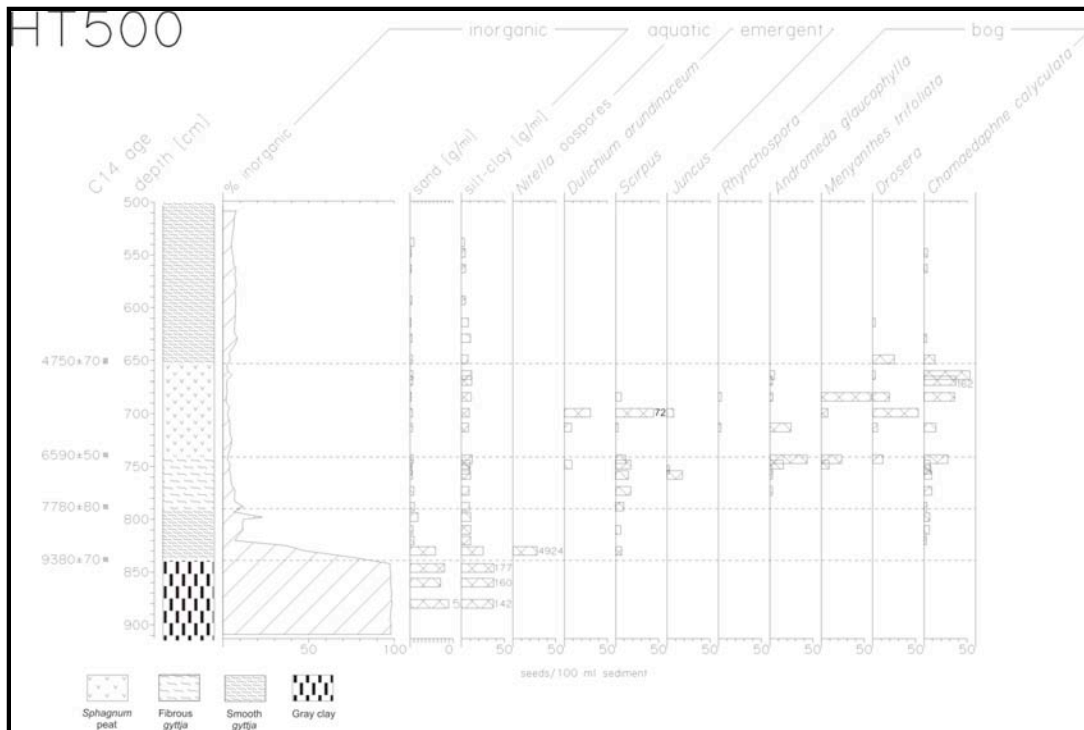


Figure. 2.2. Core HT500: sediment stratigraphy, inorganic content, and macrofossils. From Almquist *et al.* 2001.

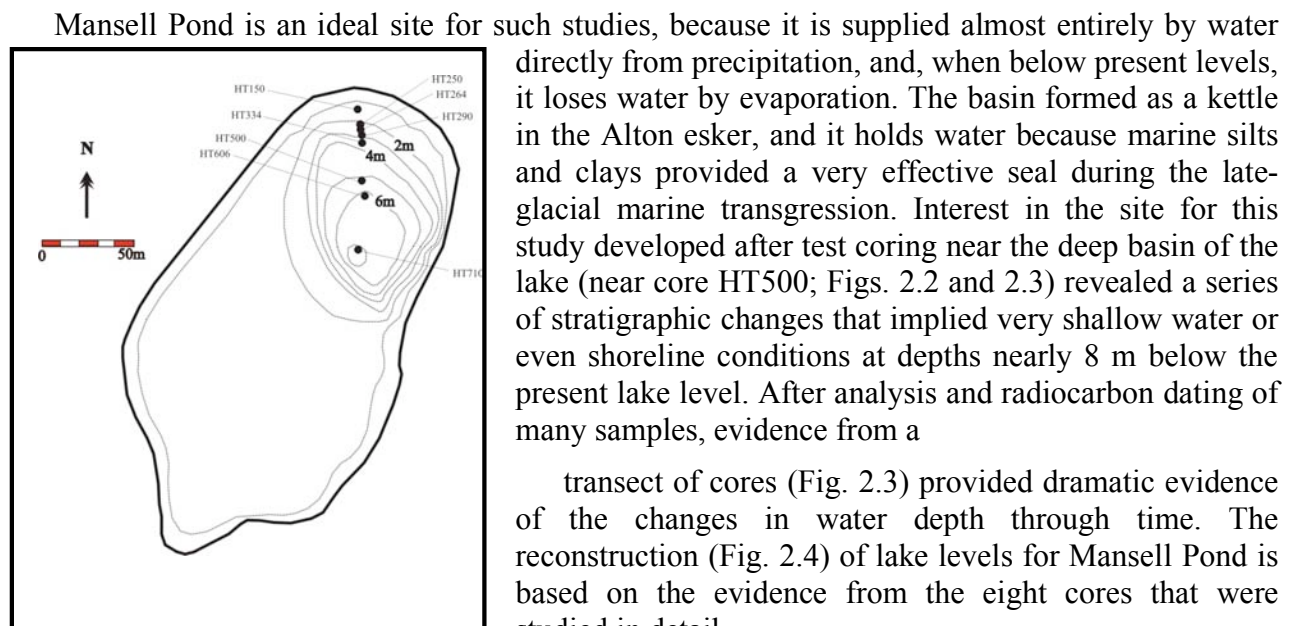


Figure. 2.3. Bathymetry of Mansell Pond and location of coring sites. From Almquist *et al.* 2001.

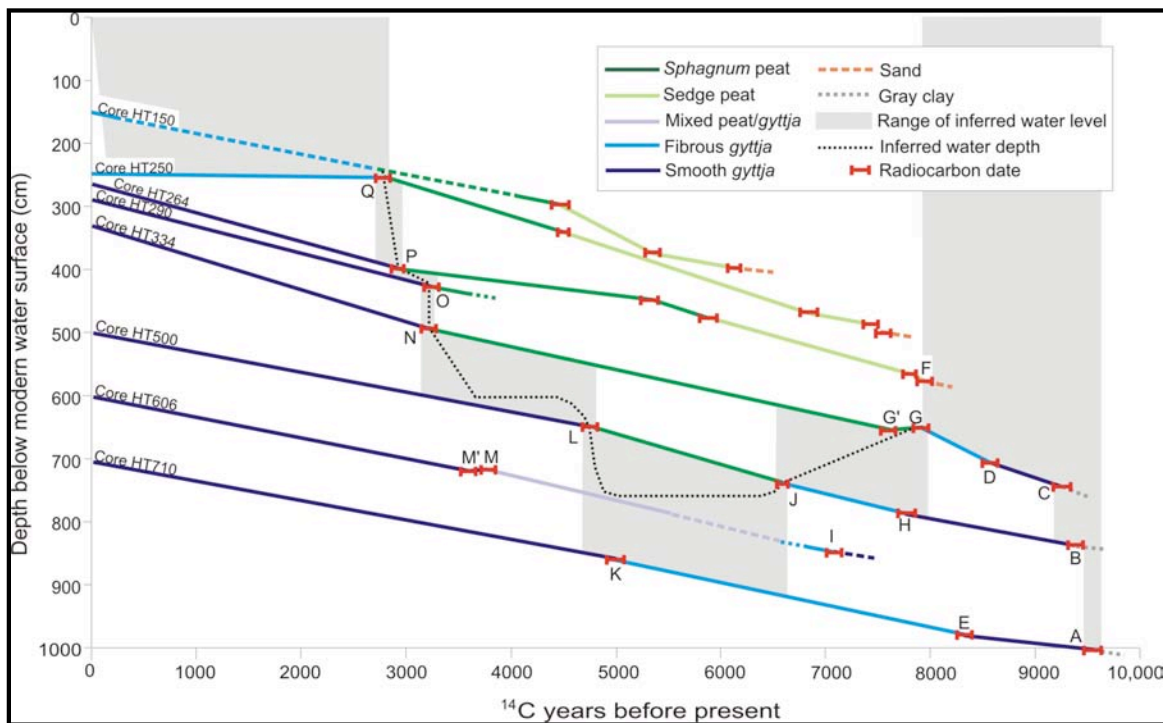


Figure 2.4. Holocene lake levels of Mansell Pond. From Almquist *et al.* (2001).

Regional similarities in paleohydrology are revealed by comparisons of the data from Mathews Pond and Whitehead Lake (Fig. 2.5) with those of Mansell Pond. The general pattern is one of a dry late-glacial, followed by rising lake levels. A relatively dry period is evident between ca. 7000 and 8000  $^{14}\text{C}$  yr BP in both northern sites. All three data sets indicate low water levels in the mid-Holocene, but the driest period is slightly earlier in Mansell Pond. At all sites the water levels have been rising consistently for the past 3000 years, consistent with the cooler, moister climate that has allowed the southern expansion of spruce populations in the region (Schauffler and Jacobson 2002).

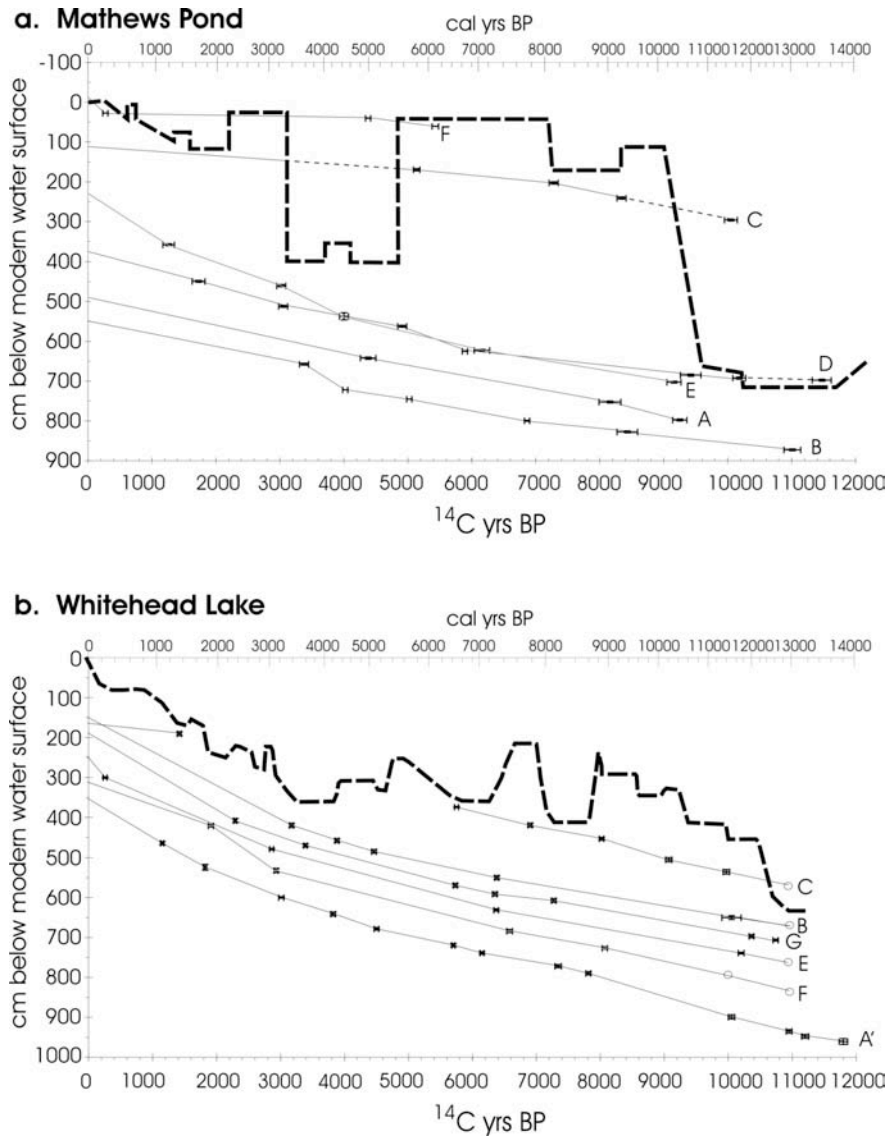


Figure 2.5. Inferred lake levels from (a) Mathews Pond and (b) Whitehead Lake. From Dieffenbacher-Krall and Nurse (2005).

## **STOP 3: PUSHAW LAKE/CARIBOU, BOG/WHITTEN BOG COMPLEX (ALICE KELLEY)**

*(nb: All dates are expressed in un-calibrated radiocarbon years BP)*

Almquist and Sanger (1995, 1999) used 31 cores, combined with radiocarbon chronology, to create a paleogeographic reconstruction of the Pushaw Lake/Caribou Bog/Whitten Bog Complex (Fig. 3.1a) through the Holocene. These reconstructions show the relative extent of water, peatlands, and cattail marshes at 10000, 9000, 8000, 6000, and 1000 BP (Fig. 3.1). Kelley (2006) estimated the amount of surface water at each time period in the lower portion of Pushaw Stream, outside the original study area, by using contour lines to extend the water bodies identified by Almquist and Sanger (1999). This provides an approximation of the changing environmental conditions in this area through time.

At 10,000 BP, an extensive lake covered much of the Pushaw area, including lower Pushaw Stream (Fig. 3.1b). Kelley (2006) interprets this lake to represent the initial phase of development of surface drainage in the area. With the withdrawal of marine waters from the Penobscot Valley circa 12,000 BP, the ground surface was mantled by a drape of glaciomarine Presumpscot silt and clay ranging in thickness from less than a meter to over 3 m. This fine-grained deposit retarded infiltration and fostered the formation of lakes and ponds. The development of newly-forming, southward-flowing drainage net-works was also interrupted circa 10,000 BP by a northwestward migrating forebulge which crossed the coastal portions of Maine prior to 10,000 BP, and is noted at Moosehead Lake, 150 km to the northwest circa 9,000 BP (Barnhardt, *et al.*, 1995; Balco *et al.*, 1998; Kelley, 2006). This localized northward tilt in an area of low-relief may have slowed drainage development, elevated lake outlets, and enhanced the growth of lakes in the region.

By 9,000 BP, open water in the Pushaw area was less widespread and separated into two discrete basins, occupying the modern day Pushaw Lake/Mud Pond/Caribou Bog and the Whitten Bog/Pushaw Stream areas, respectively (Fig. 3.1c). Local drainage returned to a generally south-flowing pattern, and drainage patterns became more integrated. At this time, a limited peatland formed in the northeastern portion of the Whitten Bog area. Open water continued to decrease with time, and by 8,000 BP peatlands had expanded in Whitten Bog and Caribou Bog, and Pushaw Lake had begun to approach its present day form (Fig. 3.1d). However, much of the southern portion of Caribou Bog remained open water, as did the eastern portion of Pushaw Stream. Cattail marshes appeared on the northeastern portion of Caribou Bog and the southwestern portion of Whitten Bog.

By 6,000 BP peatland had replaced the cattail marshes in Caribou Bog and cattail marsh filled the areas previously occupied by open water in Whitten Bog, (Fig. 3.1e). The lower Pushaw Stream area was still occupied by standing water, as was the southern portion of Caribou Bog.

Almquist and Sanger's (1999) paleogeographic reconstruction shows establishment of modern landscape conditions by 1000 BP (Fig. 3.1f). Pushaw Lake, Mud Pond, and Pushaw Stream were at approximately their current size and configuration. Peatlands had reached near modern proportions, with some cattail marshes lingering on the banks of Dead Stream near the confluence with Pushaw Stream, and north of Whitten Bog.

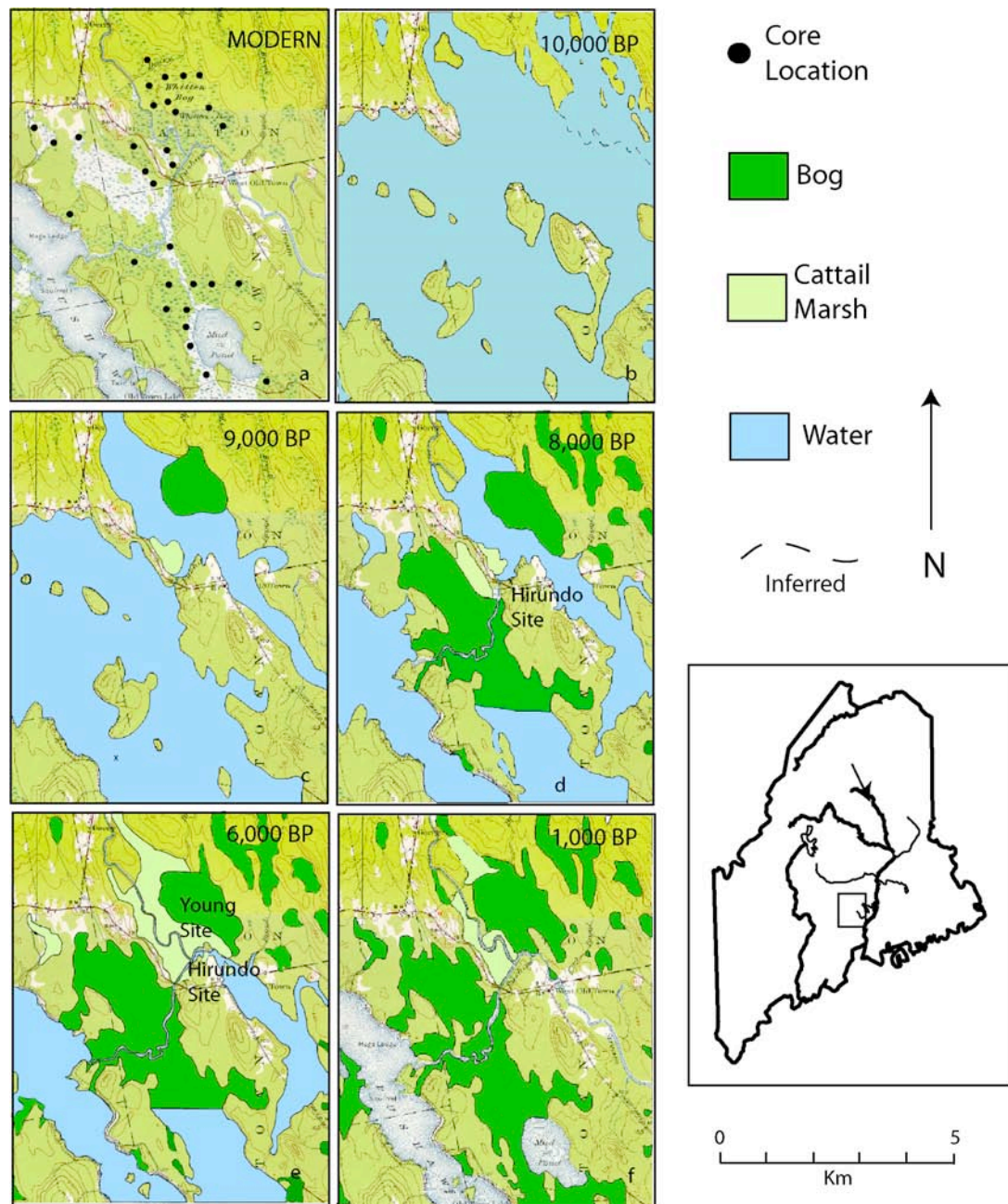


Figure 3.1 Paleogeographic reconstruction of the Pushaw Lake/Caribou Bog/Whitten Bog area. Modified from Almquist and Sanger (1999).

The nature of the confluence of Pushaw Stream with the Stillwater River is speculative at present (Fig. 3.2). Pushaw Stream cuts through a large esker before entering the Stillwater River. The timing of incision of this channel is currently unknown. Prior to that time, there may have been a lake upstream from the esker. This reconstruction, however, provides a starting point for paleogeographic reconstruction in the lower Pushaw Stream area.

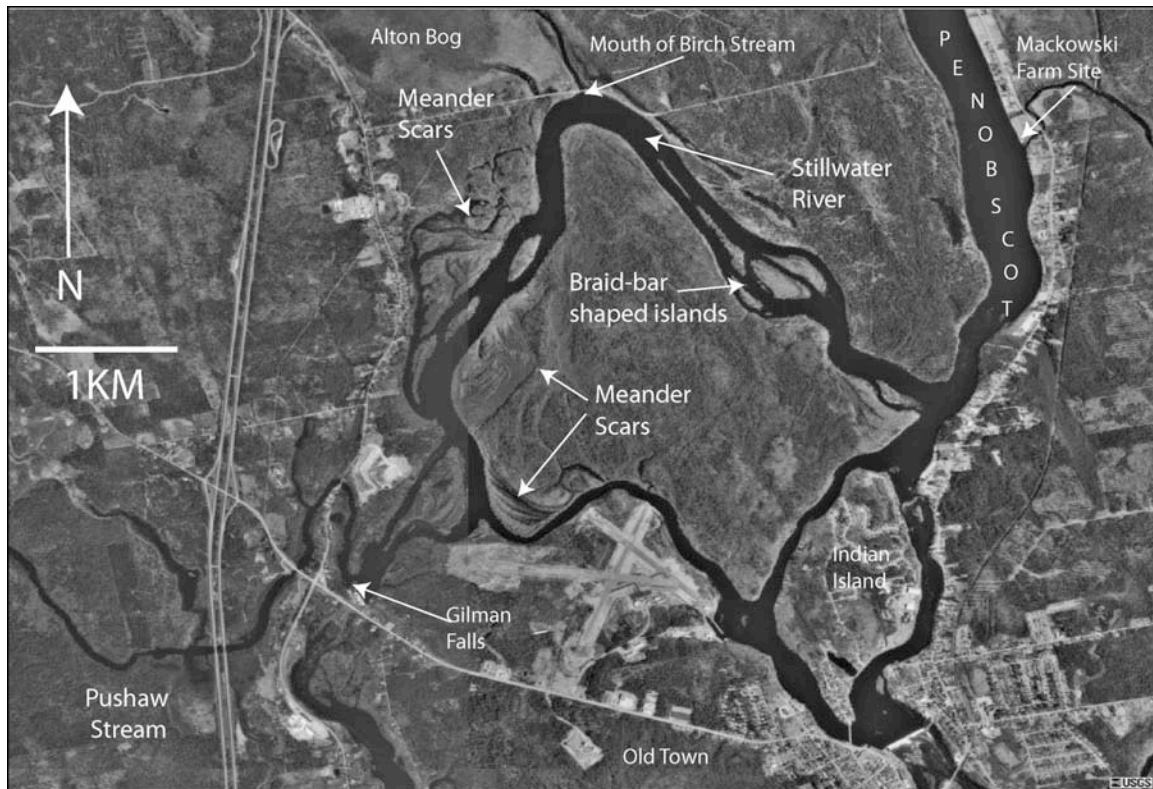


Figure 3.2 Aerial photograph of the upper portion of the Stillwater River. USGS photo

Archaeological data support these paleogeographic interpretations. The Hirundo Site, at the present-day confluence of Pushaw and Dead Streams, is located on the Early Holocene lakeshore (Fig. 3.1d). The lowest stratum of this multi-component site is dated to the Early/Middle Archaic period, circa 7,000 – 8,000 BP, on the basis of diagnostic artifacts including Neville-like points and several quartz scrapers. The Young Site (Fig. 3.1e), also near the confluence of Pushaw and Dead Streams, has evidence of Middle to Late Archaic period occupation, circa 6,000-3,000 BP, at the base of the archaeological sequence. Use of this site may have been related to the rich resources associated with the adjacent cattail marshes. The oldest evidence of occupation at sites along lower Pushaw Stream is associated with Late Archaic and Ceramic period material, circa 4,000 to 400 BP. Following Almquist and Sanger's (1999) reconstruction, this area was not available for occupation until approximately this time.

If this model is applied to Alton Bog, a beautiful large bog which we saw from the lookout point at Stop 1 (see also Fig. 3.2), the area of the bog may have been a large, shallow, Late Pleistocene/Early Holocene lake with a south-draining outlet where the present-day Birch Stream joins the Stillwater River. Northward tilt of the lake in the Early Holocene would pond water against the uplands north of the bog. The presence of these and other, similar lakes and wetlands in the region undoubtedly influenced travel and hunting routes of Late Pleistocene and Early Holocene occupants.

It is interesting to consider the possibility that the convoluted flow pattern of the Stillwater River (Fig. 3.2) may be related to the increased discharge from this lake as the south tipping limb of the forebulge moved through the region and southward flowing drainage networks were established.

## **STOP 4: PRESUMPSCOT FORMATION (*HAL BORNS*)**

We have already mentioned the Presumpscot glaciomarine clay several times. This stop will provide an opportunity to examine a nice section in outcrop.

Probably the earliest description of this clay was by Jackson (1837) who recognized the conundrum posed by marine fossils 8 m above present sea level. It was Stone (1890, p. 132), however, who seems to have been the first to make a direct connection between the clays and their glacial source. The name of the formation comes from the Presumpscot River near Portland, along which there are many good exposures (Bloom, 1960).

The Presumpscot mud is largely pulverized rock debris (“rock flour”) from the ice/bed interface. Much of the mud was deposited directly into the sea from the ice, but some may have been transported some distance in proglacial streams, and some undoubtedly came from erosion of uplands. While commonly called a clay, the composition of the Presumpscot varies from place to place. On average, it is about  $\frac{1}{3}$  very fine to fine sand,  $\frac{1}{3}$  silt, and  $\frac{1}{3}$  clay (Goldthwait, 1951; Caldwell, 1959; Borns and Hagar, 1965). It also contains “drop stones” ranging from sand grains to boulders. In the Greenbush quadrangle, Hooke has mapped a very fine sand facies, as a shallower-water deposit.

The Presumpscot may be either massive or well stratified. In the latter case, beds of silt, clay or sand can be up to several centimeters thick. In some areas, vertical desiccation cracks are observed, indicating subaerial exposure; some were found just west of the Alton esker.

Although usually not macroscopically fossiliferous, fauna reported from the Presumpscot range from forams through mammoths, and flora include seaweed, leaves, and rare tree trunks. Most common are small critters like clams (*Mya arenaria*), mussels (*Mytilus edulis*), barnacles (*Balanus balanus*), and arctic hiatella (*Hiatella arctica*). Some of these species now live only in colder water north of Maine. In the ice-proximal facies of the Presumpscot Formation, faunal species suggest arctic to sub-arctic conditions, with sea temperatures of 0 to  $-2^{\circ}\text{C}$ . Oxygen and carbon isotope ratios in the same fauna suggest water of normal salinity with a temperature between  $0^{\circ}\text{C}$  and  $+2^{\circ}\text{C}$  (Dorion *et al.*, 2001).

Landslides are common in bluffs of the Presumpscot, and have resulted in significant economic losses.

Thompson (1987) gives an extended description of the Presumpscot, from which most of the above material was derived.

Returning to the present stop, the ice margin retreated across this location  $\sim 13,500$   $^{14}\text{C}$  years ago and the sea here was then  $\sim 50$  m deep. Deep-water glaciomarine mud started to drape the landscape, including the esker here. As the sea became progressively shallower, wave action washed sediment from the esker crest down the flanks onto the deep-water clay. This process is recorded by the gradational upward coarsening of the marine sediments on the esker flank, and by the upward change in shell fossils from a few “deep” water species and numbers to many shallow water species and numbers.

## STOP 5: GILMAN FALLS (DAVE SANGER AND ALICE KELLEY)

The Gilman Falls Site (Sanger, 1996; Sanger *et al.*, 2001) is located on the north end of a bedrock-cored island at the upstream end of Gilman Falls, near the confluence of Pushaw Stream and the Stillwater River (Figs. 3.2 and 5.1). The site was excavated in 1990 - 1992, as part of the cultural resource evaluation required for the relicensing of the Milford Dam. This dam forms a large reservoir on the Penobscot River north of Old Town and Milford, the upper portion of the Stillwater River, and lower Pushaw Stream. Approximately 320 m<sup>2</sup> were excavated as part of the project, with most of the work focused on two large block excavations, 10 m x 10 m, and 20 m x 11 m.

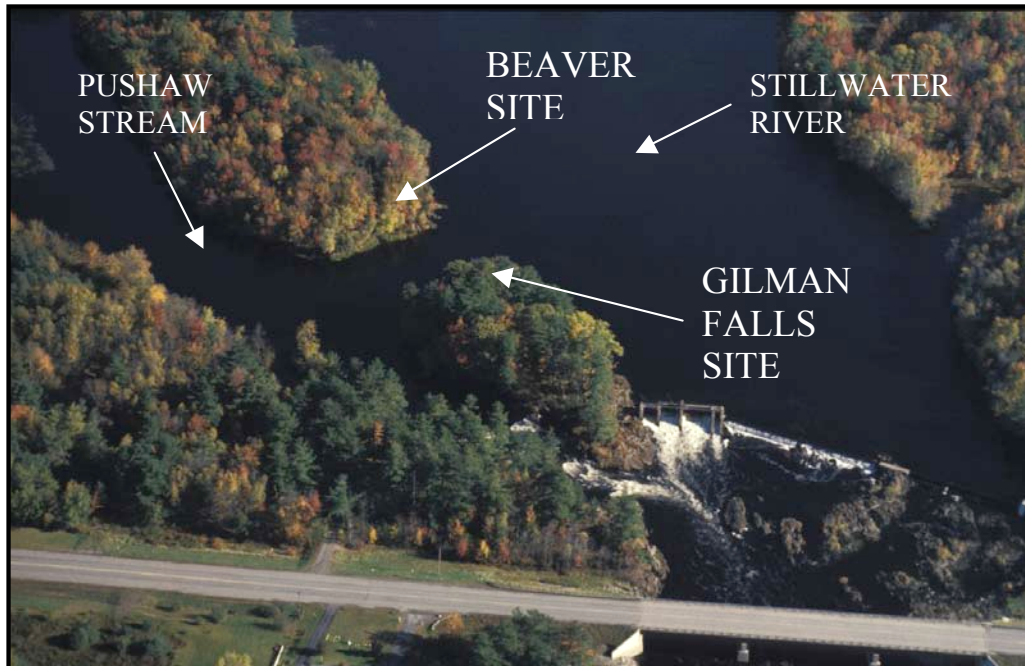


Figure 5.1: Location of Gilman Falls and Beaver Sites. Pushaw Stream enters the Stillwater River from the northwest. The falls are composed of an erosion-resistant facies of the Vassalboro Formation.

The site is composed of ~80-200 cm of predominately fine-grained alluvial sediments, either deposited on stratified sand and gravel or resting directly on metasedimentary rocks of the Kenduskeag unit of the Silurian/Ordovician Vassalboro Formation (Griffen 1976a, 1976b). Nine distinct strata are identified at the site on the basis of sediment texture and color. While some individual units vary in thickness, and pinch out within the excavated area, most strata are continuous across the site. The transition from stratified, coarse-grained material or bedrock to massive, fine-grained deposits is abrupt across the base of the entire excavation.



Figure 5.2 Stratigraphic section at Gilman Falls Site. Spodic horizons indicated by arrows.



Figure 5.3 Stone rods recovered at Gilman Falls Site

Bright to faint red, laterally extensive horizons are encountered at several elevations throughout the excavation (Fig. 5.2). These horizons occur at depths of 5–15 cm (immediately below the modern A horizon), 60–70 cm, and 100–120 cm. They range from distinct, sharply bounded layers, 5–10 cm thick, to broad diffuse bands, 15–30 cm thick. Texturally, they are identical to the surrounding sediments. On the basis of color, lateral extent, and fine-grained texture, these horizons are identified as remnant spodic horizons (Fernandez and Osher, pers. comm., 2000; Brady and Weil, 1996, p. 87–88; Callum, 1995). While not continuous across the site, they are differentiated by stratigraphic position and associated artifacts.

The Gilman Falls site has an extensive Middle Archaic component, as well as material representative of a much smaller later Archaic occupation and a limited Early to Middle Ceramic period occupation. Three stratigraphic zones are identified on the basis of stratigraphy and artifacts.

- Zone 1 encompasses the upper 20 cm of the stratigraphic sequence, shows evidence of disturbance, primarily by tree roots, and is associated with the upper spodic horizon. Archaeological material in this zone ranges in age from Late Archaic through Middle Ceramic periods, circa 4000–1350 BP.
- Zone 2 ranges in depth from 20 to 40 cm below the surface. Radiocarbon dates from this zone range from circa 3900 BP at the top to circa 6000 BP at the base. Most of the dates, however, cluster between ~4000 and 4400 BP. Zone 2 contains the middle buried spodic horizon. Artifacts representing the Laurentian and the Susquehanna traditions, associated with the Late Archaic period, are found in this zone.
- Zone 3 is ~30 cm thick, and contains a Middle Archaic (6000–4000 BP) assemblage associated with the lower buried spodic horizon. Total inorganic phosphate analyses of this horizon were substantially higher than values from other levels in the site, consistent with extensive human occupation during this time period.

Paleosols have been identified at other sites within the Penobscot Valley. At the Blackman Stream site, on the east bank of the Penobscot River near Orono, a brown to reddish brown horizon

contained charcoal dated to  $7400 \pm 140$ ,  $7760 \pm 130$ , and  $8360 \pm 150$  BP. At the Bob Site, on lower Pushaw Stream, a dark brown paleosol was associated with charcoal dated to  $3560 \pm 70$  and  $4650 \pm 70$ . At Markowski Farm, which we will visit tomorrow, charcoal associated with a dark brown paleosol returned a date of circa 6900 BP. Each of these time periods is readily correlated with periods of lower or stable lake levels noted in the paleohydrologic study of Mansell Pond (Stop 2) (Almquist *et al.*, 2001). Lowered or stable lake levels are interpreted as indicative of lower regional precipitation and river discharge. The reduced flow decreased deposition and allowed pedogenesis to take place. The lowest spodic horizon at Gilman Falls, at  $6290 \pm 160$  to  $7285 \pm 80$  BP may be correlative with the Blackman Stream and Mackowski Farm paleosols. The middle Gilman Falls spodic horizon roughly correlates with the Bob Site paleosol. The upper spodic horizon, however, represents too large a time span to be correlated with the Mansell Pond record or other paleosols in the region (Diffenbacher-Krall and Nurse, 2005).

Sanger *et al.* (2001) identify the Gilman Falls site as a quarry/workshop used for the production of stone rods (Fig. 5.3) found in burial contexts at other archaeological sites in the region. These artifacts were produced from the local bedrock. The original form was determined by the diamond-shaped cleavage pattern, which was then pecked and ground to create a rounded stone rod. Artifacts in various stages of production were found at the site.

The adjacent Beaver Site (Belcher and Sanger, 1988), is located on a peninsula at the confluence of Pushaw Stream and the Stillwater River (Figs. 3.2 and 5.1). Sediments at the site are composed of fine sands deposited on compact silt, but differ in the thickness and in number of individual units present. Gravel was exposed at the base of the deepest excavation of the western portion of the site, and may represent Late Pleistocene outwash or an Early Holocene gravel bar.

A date of  $8150 \pm 250$  BP on charcoal with associated artifacts places the earliest known occupation of the Beaver Site in the Early Archaic period, earlier than that noted at Gilman Falls. A limited Late Archaic assemblage was recovered at depth. The overwhelming amount of archaeological material at the Beaver Site, however, is from the Ceramic period. The material is near the surface, and becomes sequentially younger moving toward the bank of the Stillwater River.

Preservation at these sites is related to their location. In this region, thick sedimentary sequences have only been identified either where geologic features, such as moraines or bedrock outcrops, create local base levels, or at the mouths of tributary streams, or a combination of the two situations (Petersen and Putnam, 1992; Sanger *et al.*, 1992, 2001; Kelley and Sanger, 2003; Robinson 2001). In the former case, the geomorphic obstructions pond water during flood events, allowing fine-grained material to settle out. In the latter, sediment-rich water from tributary streams is blocked from entering the main stem of the river by hydraulic damming, and deposition of fine-grained material takes place at the mouth of the tributary. Archaeological investigations of these thick, stratified sequences in the central Penobscot Valley Sanger *et al.*, 1992, 2001; Robinson, 2001) and along the Piscataquis River (Petersen and Putnam, 1992) have established that humans occupied the region during the Early and Middle Archaic period. This interval was once thought to be a time of little interior occupation in northern New England due to the low carrying capacity of the boreal forest (Fitting, 1968; Ritchie, 1971).

## **STOP 6: MACKOWSKI FARM SITE (*BRIAN ROBINSON*)**

### **Sunkhaze Stream Locality**

Two archaeological sites have been studied near the mouth of Sunkhaze Stream a tributary to the Penobscot River. Although located on opposite sides of the mouth of the stream, the sites are in different geological settings and represent different cultural activities. The Sunkhaze Ridge site (“1” in Fig. 6.1) was excavated in 1922 and 1923. It represents the first Middle Archaic period cemetery recognized in Maine, and is dated to approximately 7000 radiocarbon years ago. It is situated on a Late Pleistocene/Early Holocene feature deposited during higher water levels. Burial activities occurred long after the land surface had formed. The Mackowski Farm site (“2” in Fig. 6.1) is a deeply stratified occupation site north of the outlet of Sunkhaze Stream. Here there are multiple occupation floors within the fluvially deposited sequence. At this site, the chronology of the archaeological remains contributes to the interpretation of the depositional sequence.

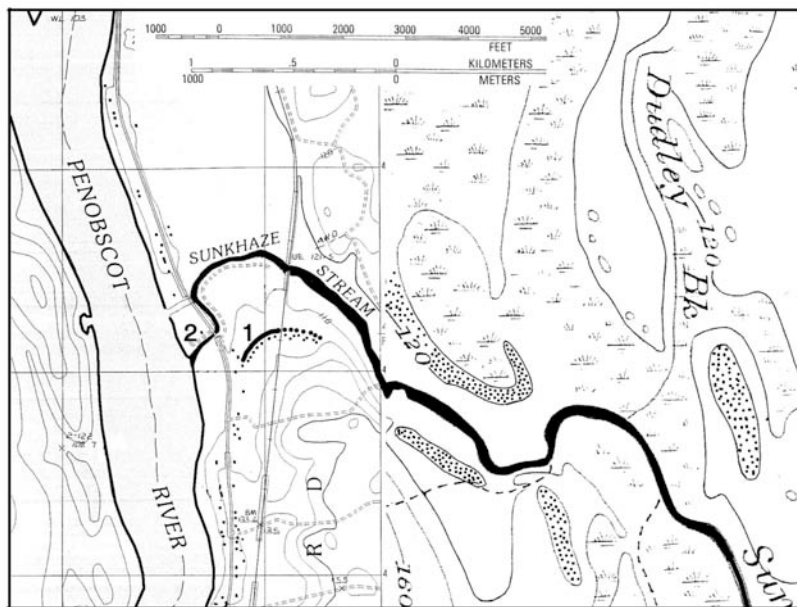


Figure 6.1 Sunkhaze Stream Locality. Locations of Sunkhaze Ridge and the Mackowski Farm sites are indicated by 1 and 2, respectively. Landforms composed primarily of sand are indicated by stippling.

### **The Sunkhaze Ridge Site**

The Sunkhaze Ridge Site (74-10) is located on a curved sand ridge, or berm, deposited on the glaciomarine Presumscot Formation and, at its western end, on lag boulders. The site was discovered and excavated in 1922 and 1923 by a civil engineer, Herbert A. Harnden, when ~200 meters of the ridge were removed for road fill during repair of a flood-damaged bridge. Harnden wrote a brief, but remarkable, report on the site that was provided to Robinson by his son, Everel Harnden (Robinson, 1992). Interviews in 1986 with John Costigan, a local resident who watched the excavations as a young boy, indicated that the graves came from the most prominent bend in the ridge (Robinson 2001, p. 151). When investigated in 1987, the banks of the sand pit operation were still visible and the location of the berm (Fig. 6.1) was reconstructed from aerial photographs. Everel Harnden participated in additional excavations by Warren K. Moorehead in 1923. Photographs were taken of the site during this work (Fig. 6.2).



Figure 6.2 Photograph of the Sunkhaze Ridge Site taken in 1923. The photograph shows four to five feet of sand deposited on top of the dense clay basement. Robert S. Peabody Museum

The Sunkhaze Ridge berm is part of an extensive system of elongate, curved sand deposits that extend up Sunkhaze Stream (Fig. 6.1). It is postulated that they are related to Late Pleistocene or Early Holocene fluvial, lacustrine or eolian environments. Currently, there is no evidence of occupation associated with the period of formation of the berms. Rather, the high sandy location was chosen as a burial place thousands of years after it was formed.

### **Mackowski Farm Site**

The Mackowski Farm (74-14) site is located on the upstream bank of the confluence of Sunkhaze Stream and the Penobscot River. It is upstream of the Sunkhaze Rips (rapids), mapped by Treat (1820), but now inundated by the headpond of the Milford Reservoir. Archaeological material has been collected at Mackowski Farm since the early 1900's. As the site of amateur collecting, it produced a large number of artifacts dating from the Middle Archaic period to the Contact period (including a lead baling seal and glass beads). Top soil stripping operations removed the upper meter or more of sediment from most of the site in the 1970's. It was first professionally excavated in 1987 (Robinson 2001), with additional work in 2003 and 2005.

The Makowski Farm site is composed of a thick, 2+ m, sequence of fine-grained sediments resting on a boulder/gravel lag (Fig. 6.3). Strata descriptions are taken from field notes and profiles completed in August of 2005. Individual strati-graphic units were identified on the basis of sediment texture and color. The upper 1+ m of the stratigraphic sequence is recorded from archaeological excavations. The lower 1+m is the result of bucket auger testing from the floor of an excavation unit.

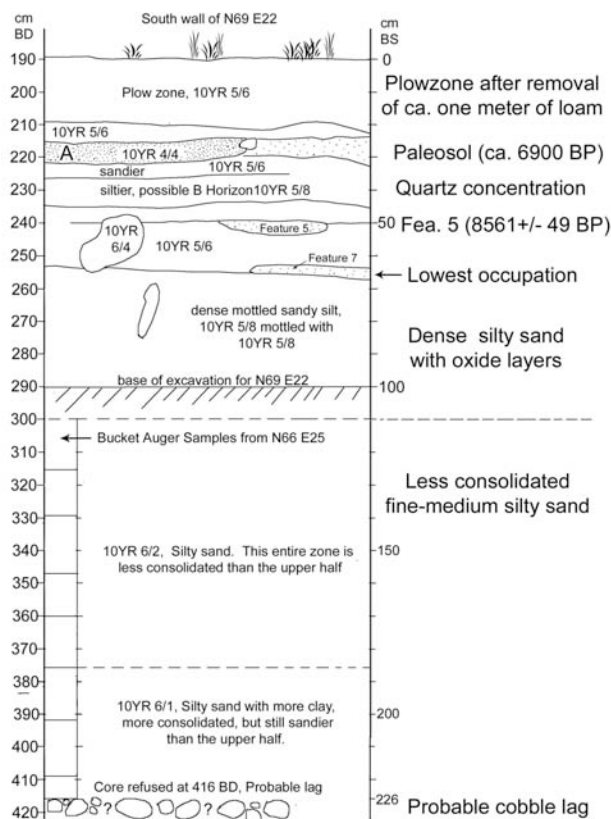


Figure 6.3. Stratigraphic section at the Mackowski Farm site.

feature was AMS dated to 8561±49 B.P. Below this there is at least one more occupation level with a high density of flaking debris and features. It is thought that the lowest levels represent relatively thin and rapid depositional episodes that helped to preserve the fragile burned fish bone.

The 6900 year old paleosol appears to best represent the age of the cemetery at the Sunkhaze Ridge site. Grave goods recovered from the cemetery were dominated by full-channelled gouges and stone rods, similar in form and lithology to those recovered at the Gilman Falls Site. The paleosol at the Mackowski Farm site produced manufacturing debris from both of these artifact forms. This Middle Archaic period date correlates well with the lower cultural horizon at the Gilman Falls site, which produced many fragments of both artifact forms.

One well developed paleosol that was traced across the site dated to ~6900 <sup>14</sup>C years BP. A series of occupation levels were encountered below this level. Although far less visible, they had tightly clustered vertical distributions of artifacts, flaking debris, and hearths. Visible stratigraphy is largely limited to anthropogenic soil changes, with discontinuous layers. The two lowest levels also contained a series of burned hearths with an abundance of burned bone, mostly fish (Fig. 6.4). Feature 5 contained thousands of fish bones and a number of charred hawthorne nuts or seeds. A single hawthorne seed from this

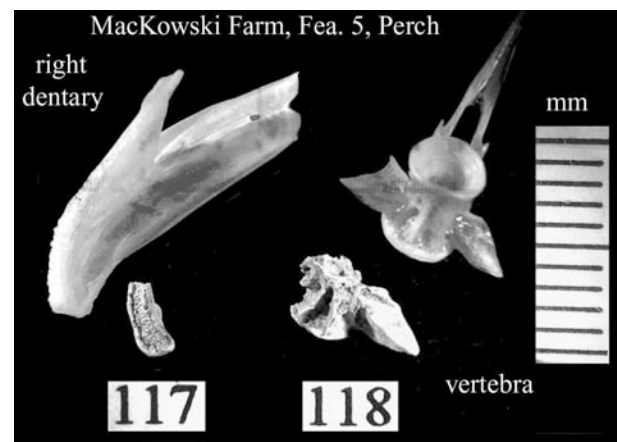


Figure 6.4 Fish bones recovered from the Mackowski Farm Site, circa 500 <sup>14</sup>C years BP.

Of considerable interest is a concentration of quartz scrapers and flaking debris below the paleosol, possibly associated with a buried B Horizon. Quartz-dominated assemblages are characteristic of the Early Archaic period from southern New England to the Kennebec River, but are less common in the Penobscot drainage. Below the quartz, the level of the hearth dated to 8561 BP is dominated by felsite tools including large side-scrapers and cores, but so far lacking ground stone tool fragments. Finally, the deepest occupation level produced large quantities of felsite, as well as a large core of Munsungun chert and numerous chert flakes. Thus far there is no evidence of biface technology in the Early Archaic period levels; this is characteristic of the Gulf of Maine Archaic tradition (Robinson, 1992). The chert level may represent an intermediate technology between the Early Archaic and Late Paleoindian periods, but more work is required to resolve this issue.

### *Geoarchaeology of the Mackowsky Farm Site*

The well-stratified sequence at the Mackowski Farm site provides important information relative to the Early Holocene geologic development of the Penobscot River. The boulder gravel at the base of the section can be observed in the bed of Sunkhaze stream during low water, and is interpreted to be a lag deposit resulting from erosion of till. The abrupt change from this lag to the overlying sands is indicative of a change from erosion to deposition in the sedimentary regime. This change may have been related to the establishment of a local base level at the Milford/Old Town falls or Sunkhaze Rips downstream or to development of a similar local base level upstream, of both.

Deposition of most of the remaining strata at the Mackowski Farm site is attributed to alluvial processes, primarily during spring floods. These sediments enclose archaeo-logical remains that are attributed to short-term occupations, and thus provide a time marker in sedimentary record. The exception to this style of deposition is the upper, darker brown horizon, encountered just beneath the plow zone, which is correlated with the Middle Archaic occupation at the Sunkhaze Ridge site. Formation of the paleosol may be related to reduced sedimentation brought about by a period of lower spring runoff. The Middle Archaic age of this feature is roughly correlative with the lower paleosol noted at the Gilman Falls (Sanger *et al.*, 2001) and with one of the dry periods recognized at Mansell Pond.

The archaeological material at the Mackowski site is indicative of short, but repeated visits of people to the area, perhaps to take advantage of the floral and faunal resources that existed at the mouth of Sunkhaze Stream and on nearby islands in the Penobscot River. Due to the proximity of the two sites, it is tempting to make an association between the Middle Archaic occupation at Mackowski and the mortuary site at Sunkhaze Ridge. More detailed analysis of archaeological sites in the region may help address this issue.

## STOP 7: STRATIGRAPHY IN 6 M TERRACE (*ROGER HOOKE*)

### Penobscot River Terraces

As mentioned earlier, terrace remnants occur at elevations of ~ 2, 4, 6, and 10 m above the present floodplain along the Headwaters and Island Divisions of the Penobscot (Figs. 0.3 and 7.1). My working hypothesis is that the sands and gravels underlying these terraces were deposited as subaerial outwash trains or braidplains extending downstream from the retreating glacier and graded to sea level. During a period of more rapidly falling sea level, an earlier braidplain would be incised (and left as a terrace). During a subsequent period of stable or slowly falling sea level, a new braidplain would form within and downvalley from the earlier one. The terrace sequence would thus telescope out downvalley. Borns and Hagar (1965) have proposed a similar history for terraces in the valley of the Kennebec River in western Maine.

Although terraces can be found, intermittently, at most or all of the above levels along the entire 100 km stretch of the river north of the Greenbush quadrangle, this is probably not a valid correlation tool. As downcutting progressed, the river encountered the bedrock sills and other obstructions that now form the riffles discussed earlier (Figs. 0.4 and 7.1). These riffles would have decoupled the river upstream of the sill from any effect that was propagating up from lower reaches. Thus, some terraces, when traced upstream, will probably end at such riffles, and higher terraces downstream may correlate with lower ones upstream.

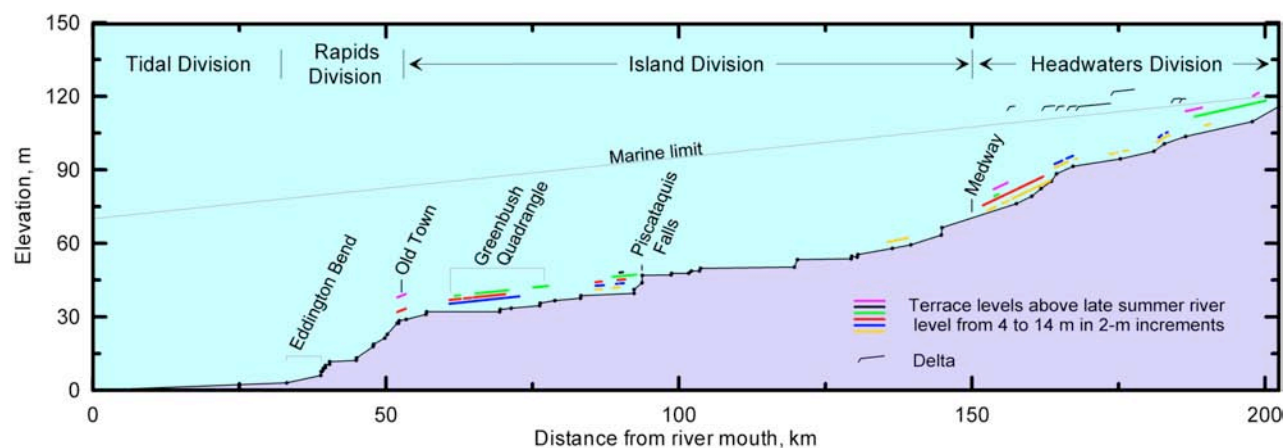


Figure 7.1 Longitudinal profile of the Penobscot River showing terraces and bedrock sills resulting in rapids or falls. Vertical elevation of terraces above river is exaggerated.

Mapping to date in the northeast quadrant of the Greenbush quadrangle (Fig. 7.2) has established that a braidplain migrated over and buried Presumpscot silts and clays, which in turn rest on till or bedrock. Locally, the Presumpscot is buried by pebble gravel, but that appears to be primarily in a place where the depositing river was simultaneously eroding part of an esker and acquiring the gravel therefrom. Elsewhere, the braidplain sediment consists of stratified sands with occasional beds of fine gravel. The unit is ~9 m thick 4 km south of the quadrangle boundary and its surface is our 6 m terrace (6 m above the present floodplain). The surface drops (and the unit thins) over the next kilometer southward but it is not yet clear to what extent this is due to a gradual feathering out of the original unit, to subsequent erosion, or to the telescoping effect. Some linear deposits of granule gravel are also present; these may be channel beds. In the far northeastern corner

of the Greenbush quadrangle the 6 m terrace has been dissected, leaving terraces also at ~2 and 4 m above the present floodplain. A paleochannel remnant is also present in the 6 m surface there.

Symbol	Description
Ha	Modern floodplain. About 4 m above late-summer river level
Hw	Holocene wetlands. Dark gray or black organic muck.
Qe	Sand dunes. Very fine to medium sand, moderately to well sorted, commonly in dune forms.
Qf	Medium sand up to pebble gravel deposited in a fan shape downflow from an esker head (Qger). Fans are inferred to be formed in a submarine environment downflow from a portal in which an esker head was formed.
Qge	Esker gravels. Ridges of pebble/small cobble gravel.
Qger	Esker head. In downflow direction, esker rises gradually and then drops off sharply or (in some quadrangles) merges with a delta.
Qp	Presumpscot Formation. Well sorted glaciomarine silt, clayey silt, or clay. Commonly light gray (N6? or N7?), but may be light olive gray (5Y5/2). Pale yellow brown (10Y6/2) where oxidized.
Qpns	Nearshore Presumpscot. Well sorted fine to very-fine sand, commonly pale yellow brown. Inferred to be deposited in a marine environment with slightly higher energy than the Qp facies. Commonly found between Qp and higher ground.
Qsp <sub>x</sub>	Braidplain. Medium sand or larger material, up to pebble gravel. Commonly well stratified. Inferred to be a Late Pleistocene - Early Holocene braided river deposit formed as the glacier retreated and sea level fell. The river remained graded to sea level, resulting in a series of terraces. The digit at the end of the symbol is the height, in meters, above the modern floodplain.
Qt	Glacial till. Extremely poorly sorted silty fine to medium sand with granules and small pebbles in soil probe samples, and pebbles, cobbles, and boulders in road cuts. Ground surface commonly has scattered cobbles and boulders.
Qtw	Water-washed till. In the Greenbush Quadrangle, this is the inferred origin of a gravel deposit in the SW corner of the quadrangle. (Not on this map).
af	Artificial fill emplaced by humans
brk	Bedrock. (Size commonly exaggerated to make visible on map.)

\*Pairs of thick red lines border opposite sides of inferred paleochannels

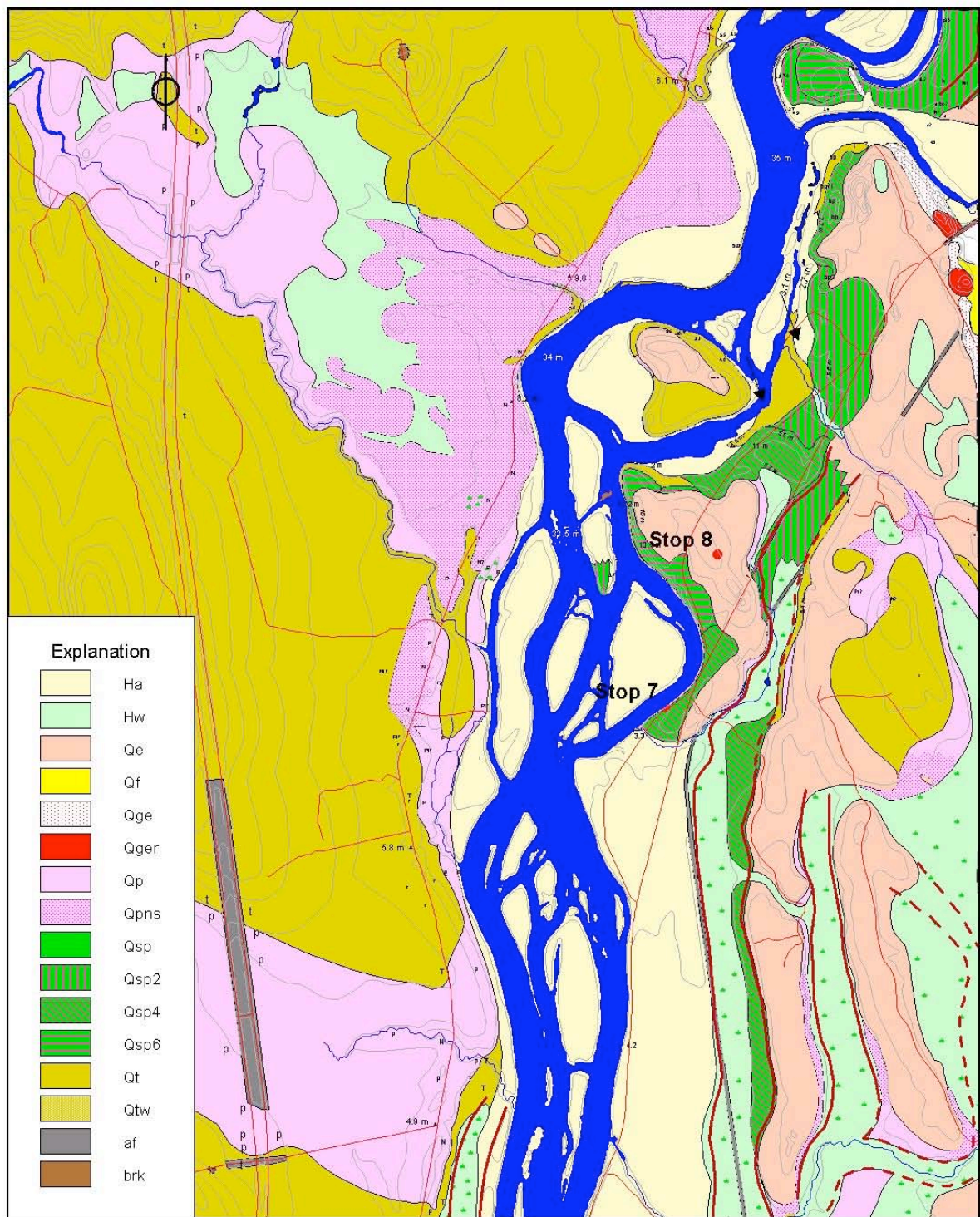
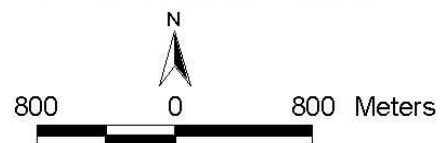


Figure 7.2 Preliminary surficial geologic map of the NE quadrant of the Greenbush Quadrangle

Mapped by Roger LeB. Hooke,  
Robin Wiesner, and Elisabet Metcalfe, 2005-2006



At this stop we will look at an exposure of the braidplain sediment and underlying Presumpscot clays in the 4 m terrace.

Braidplain pebble gravels are also present at least 20 km south of the Greenbush quadrangle, but there they do not form terraces. Rather they are overlain by floodplain sands and silts. During the archeological excavations at Gilman Falls (Stop 5), a date of ~7.8 (cal) ka was obtained from a buried soil at a depth of ~1.5 m in such floodplain sediment. Another date of ~8.9 (cal) ka was obtained from a depth of ~1 m in the floodplain near Blackman Stream, 1 km east of Orono (Kelley and Sanger, 2003). In both instances, the dates were on charcoal in hearths associated with artifacts. Also in both instances the floodplain deposits were just upstream from bedrock sills that had stopped the downcutting. It seems likely that the change to fine-grained sedimentation reflects the effects of exposing these sills (Kelley and Sanger, 2003, p. 125), and perhaps also of exposing the next sill upstream as that would both trap gravel upstream and lead to rapid lowering of the river gradient between the two sills. Thus, the dates probably reflect the time that the sills were encountered. The transition from the gravels to fine-grained sedimentation is estimated to have occurred at ~10 (cal) ka. Thus, all of the downcutting must have occurred between ~14 and ~10 (cal) ka, with slow aggradation since that time. It is likely that much of the downcutting occurred in the earlier part of this time interval, particularly in the northern part of the drainage where the effect of falling sea level would have been felt first.

An additional factor that may have influenced the transition from gravel to flood-plain deposition is the decrease in discharge in the Penobscot River at about this time [~11 (cal) ka], when tilting during isostatic rebound shifted the drainage from Moosehead Lake into the Kennebec River (Kelley and Sanger, 2003). Other drainage changes took place in northern Maine as the ice withdrew, opening outlets to the north and thus also diverting water from the Penobscot drainage (D. Putnam, oral comm., 11/05).

## **STOP 8: DUNES (*ROGER HOOKE*)**

Further south along the eastern edge of the Greenbush quadrangle there are linear wetlands, interpreted to be paleochannels (Fig. 7.2). A granule gravel is encountered along the eastern edge of one such wetland. Topographic terraces capped by fine to medium sand border the channels. Dune forms are common, suggesting that the sand is all eolian. The sand rests either on till or on Presumpscot clays, not on stratified sands or gravels. Thus, the channels appear to have developed on a surface of low relief on till which had been extensively draped by Presumpscot silts and clays. The channels were presumably the source of the sand. The distribution of the sand with respect to the channels suggests that prevailing winds were westerly.

At this stop we will look at some dune forms that do, however, rest on stratified sands.

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# ROAD LOG

Saturday, June 3

- 0.0 Edith Patch Parking Lot Parking turn out on Long Road. Turn left out of lot. Jog right at 0.4 mi and left at 0.5 mi (outside the field house).
- 0.7 Turn right on College Ave. At the traffic light at 1.7 miles keep right and go straight through the intersection
- 4.0 Turn left on Gilman Falls Ave
- 4.4 Flashing light. Remember this point. You will be sent back to it three times. Turn right this time, on Route 16.
- 4.9 Cross Pushaw Stream. See Figure 3.2 in Guidebook
- 7.4 Cross I-95
- 12.4 STOP 1a – Turn right into entrance to Sargent Pit. After pit stop, return to vehicles and head back the way you came - south
- 13.2 STOP 1b Turn left into entrance to Engstrom pit. After stop return to vehicles and continue south.
- 14.1 STOP 2 Mansell Pond. Turn right into gravel pit and drive down to right. After stop, return to vehicles and return to the flashing light at the Route 16/Gilman Falls Ave. intersection. Reset your odometer at the flashing light.

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- 0.0 Flashing light. Continue south on Route 16.
- 2.6 Turn right on Kirkland Rd
- 4.3 Bear left on Poplar Rd
- 7.7 STOP 3. Turn right into Sewell Park Picnic Area. Lunch. Then Pushaw Bog stop. After stop, return to flashing light and reset odometer.

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- 0.0 Flashing light. Continue north on Route 16.
- 2.3 STOP 4. Turn left into town of Old Town gravel pit. The turn is a dozen yards or so before you come to the Route 116 intersection. After stop, return to, you guessed it, the flashing light and reset odometer.

-----

- 0.0 Flashing light. Turn left on Gilman Falls Ave
- 0.1 STOP 5. Gilman Falls archeological site. The bus may turn in and stop at the head of the gravel road. In that case, following vehicles should park on the right side of the highway. Alternatively, the bus may go down the gravel road or may continue across the bridge and turn on a paved road. In either of these cases, follow the bus.

After the stop, return to campus, following our outbound route in reverse. (That means continuing east on Gilman Falls Ave and turning south on College Ave., about 0.3 miles after the bridge.)

## Sunday, June 4

- 0.0 Edith Patch Parking Lot. Turn right out of lot.
- 0.1 Turn right on Rangeley Rd.
- 0.8 Turn left on Park Street (Route 2)
- 4.0 Turn right at stop light in Old Town and cross river. Follow Route 2 as it bears around to the left through Milford and heads north.
- 7.7 STOP 6 Mackowski Farm archeological site. Driveway is on left immediately after crossing Sunkhaze Stream. After stop, return to Route 2. Reset odometer.

-----

- 0.0 Turn left on Route 2, continuing north.
- 5.8 Turn left on Lower Falls Road
- 6.8 STOP 7: Braidplain section. After stop, continue north on Lower Falls Rd
- 6.9 Turn left on Route 2
- 7.5 Turn left on gravel road just before Quality Carbide sign
- 7.8 STOP 8: Dunes. **Please do not walk on the recently seeded areas.**  
After stop, return to vehicles and to Route 2.
- 8.05 Turn right on Route 2
- 10.2 Turn right into boat landing. *LUNCH*. End of trip. After lunch, return to campus following our outbound route in reverse.

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