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The cover photograph is by Arthur M. Hussey II, to whom this guidebook is dedicated. Arthur Hussey was an accomplished photographer and his numerous photo collections highlighted many aspects of the natural beauty of southwestern Maine. The photo was taken by Arthur at a location about a kilometer south of Lookout Point along the western shore of Harpswell Neck. Arthur first began mapping in this area in 1962, and his 1965 NEIGC field trip visited exposures nearby. The view in the photo is towards the south and the exposures are east-dipping metamorphosed Ordovician volcanic rocks of the Cushing Formation. Arthur's hammer for scale.

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GLACIAL DEPOSITS, SUBGLACIAL MELTWATER PROCESSES, AND DEGLACIATION OF THE BELFAST AREA, MIDCOAST MAINE: NEW INSIGHTS USING LIDAR IMAGERY

By

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INTRODUCTION

Much has been learned about glaciation, deglaciation, and related sea-level change in coastal Maine after decades of mapping and research. During the last few years, however, LiDAR imagery has revealed “new” glacial features in unprecedented detail. On this trip we will visit moraines, glaciomarine deltas, and paleo shorelines formed in a marine environment during the late-glacial highstand of relative sea level. LiDAR images and recent MGS mapping of these and other features will be examined to see how they relate to deglaciation of the Penobscot Bay region and changing sea level. Was Penobscot Bay occupied by a lobe of ice, or by a calving bay? Then we’ll visit good examples of large-scale subglacial meltwater scour zones and esker networks that were unknown prior to the availability of LiDAR.

LiDAR is short for Light Detection And Ranging. The raw data used to produce lidar imagery are acquired by airplanes equipped to shoot multiple laser beams at the landscape while flying overhead. The laser beams are reflected back to the plane and the vast number of recorded arrival times yield numerical data that can be processed to reveal images of either the ground surface or features such as buildings and tree cover. When the returns are processed to yield “bare earth” imagery the result is especially useful to geologists because it renders the forest cover invisible and shows Earth’s surface in startling detail. Shorelines, eskers, and thousands of subtle moraine ridges – some of them only 1–2 m high – are among the glacial-age features that have been revealed by LiDAR imagery of coastal Maine.

At those stops where LiDAR is discussed, the imagery will be examined from the following perspectives: (1) What Pleistocene deposits and erosional features have been newly revealed by LiDAR? (2) How do these landforms compare with those previously identified during surficial geologic mapping? (3) In what ways does LiDAR inform our understanding of regional behavior of the Laurentide Ice Sheet in southern Maine? (4) Does evidence of these processes support or refute what might be expected on theoretical grounds and modeling of the ice sheet?

REGIONAL OVERVIEW AND PREVIOUS WORK

The area visited during this trip is west of Penobscot Bay in Maine. It straddles the border between an area along the coast that was submerged by the sea in late-glacial time, and hilly uplands to the north that lay above the marine limit. The marine submergence resulted from isostatic crustal depression during and lingering after recession of the Laurentide Ice Sheet. Consequently the margin of the Laurentide Ice Sheet was in contact with the sea in lowland areas during deglaciation of southern Maine. Information from glaciomarine deltas shows that they were deposited in shallow waters, generally less than 100 m deep, and thus they are interpreted to have formed along a grounded tidewater glacier margin (Thompson et al., 1989).

There have been many studies of glaciomarine sediments and depositional environments in Maine, including moraines, subaqueous fans, and deltas. Much of this work was stimulated by reconnaissance surficial geologic mapping and related research projects by the Maine Geological Survey (MGS) in the 1970s and 1980s. Detailed mapping has occurred in more recent years and continues today under the MGS-USGS STATEMAP cooperative program. Faculty and student research at various universities has spun off from the mapping projects, leading to characterization of Maine’s glaciomarine deposits and development of integrated facies models (e.g. Smith, 1982; Smith and Hunter, 1989). Researchers have demonstrated the importance of sedimentation processes along the grounding line of the ice sheet in producing moraines and other ice-marginal deposits below the marine limit.
Ice sheet thermal regime and subglacial drainage

While an ice sheet may be frozen to its bed, the bed of the Laurentide Ice Sheet in Maine was, for the most part, at the pressure melting point and liquid water was present. Under these conditions, geothermal heat reaching the bed from Earth’s interior and frictional heat produced near the bed by shear in the basal ice may exceed conduction of heat to the glacier surface. This results in melting of basal ice. Along a flowline extending outward from the center of an ice sheet, there is likely to be a zone of such melting near the center followed by a zone in which cold ice, advected downward from the surface, results in freezing at the bed. Further out along the flowline, beginning roughly at the equilibrium line, frictional heating re-initiates melting (Hooke, 2005, p. 135-137).

Where melting first starts in this latter zone, the melt rate is low and the water may be lost downward to groundwater, which then flows outward and re-emerges in springs in the glacier forefield. Groundwater flow rates are low, however, so once the capacity of this flow system is exceeded and water pressure at the bed approaches the ice overburden pressure, water begins to accumulate and to flow outward along the bed. Initially it may flow as a thin film, but as more water accumulates it will collect into broad low conduits (Fig. 1).

Which way will subglacial water flow?

Water flow on land is downhill. It is normal to topographic contours, which are contours of equal hydraulic potential, \( \Phi = P_w + \rho gz \). \( P_w \) is the water pressure, \( \rho \) is the density of water, and \( z \) is the elevation above a datum. Subglacial water flow is likewise normal to contours of equal hydraulic potential. On a subglacial landscape, however, these contours – the dashed lines in Figure 2a – are lines of intersection between equipotential surfaces within the ice and the subglacial topography. The surfaces of equal hydraulic potential within the ice dip upglacier at roughly 11 times the slope of the glacier surface (Fig. 2b) (Shreve, 1972, 1985). Thus, as one might expect, subglacial water flows in a direction determined in part by the subglacial topography and in part by the slope of the glacier surface.

Formation of eskers

In Figure 2a, the solid contours depict a ridge oriented normal to glacier flow. The water flow shown on the subaerial landscape is normal to these contours and thus toward the ridge, and then southward, parallel to the ridge. However, when this topography is submerged beneath an ice sheet (Fig. 2b), the equipotential contours in the subglacial environment may guide the water toward the ridge and then over it at its lowest point as shown. Note (Fig. 2b) that owing to the dynamic pressure of the ice against the stoss side of the ridge, the equipotential contours are closer together over the ridge. Just as water flows faster on steeper subaerial slopes, subglacial water should flow faster through passes in ridges. Owing to the extra sediment transport capacity thus provided, we may expect
sedimentation to be reduced through the pass, and indeed eskers are commonly observed to be discontinuous through such passes (as shown in Fig. 2a).

Figure 2. (a) Contour map of a landscape on which are superimposed contours (dashed) of equipotential from a time when an ice sheet covered the landscape. (b) Topographic cross section from a time when ice was present showing the equipotential surfaces in the ice sheet. Numbers are arbitrary units. Owing to the pressure exerted by the ice against the ridge, the distance required for a 10 unit pressure drop is longer between B and C than between A and B (Modified from Hooke, 2005, Fig. 8.24).

For an esker to form, small conduits like those shown in Figure 1 must grow and coalesce to form a much larger conduit. Energy dissipated by the flowing water melts ice and thus enlarges conduits. Melting cannot occur, however, if the temperature gradient in the ice, with temperatures becoming colder upward, is so steep as to conduct all of the dissipated energy (heat) upward, rather than leave it to act on the walls of the conduit (Hooke and Fastook, 2007). Numerical modeling of the advance and retreat of the Laurentide Ice Sheet over Maine suggests that it was only within 5 to 10 km of the ice margin that the temperature gradient was low enough to permit enlargement of conduits by melting.

Once a conduit is able to enlarge itself by melting, it is likely to become sharply arched as this is the most stable form (Shreve, 1972). Then, the equilibrium situation toward which the system tends to evolve, is one in which the melt rate on tunnel walls equals the rate of tunnel closure due to the weight of the overlying ice. The driving force for this closure is a small difference in pressure between the overburden pressure and the water pressure in the conduit. As the tunnel walls melt, sediment in the ice is released into the water. This can overload a stream in a short distance, initiating deposition of an esker. We think the stream building the esker is small compared with the size of the esker (Fig. 3), and is held on top of the esker by tunnel closure rates that are higher near the base of the esker.
Where the ice is thin, near the ice sheet margin, tunnel closure rates are low and are unable to keep the stream on top of the esker. Here, streams commonly slide off the esker on one side or the other (Fig. 3b) and start forming a subsidiary esker subparallel to the parent (Fig. 4). Stone (1899) referred to these sections of eskers as reticulated; Hooke (2005, p. 240f) calls them esker nets (See Fig. 13).

**Figure 3.** a. Subglacial stream held on top of an esker by the higher closure rate of ice at the base of the esker. (Modified from Hooke, 2005, Fig. 8.28). b. Esker in mouth of conduit in Aktineq Glacier, Nunavut, Canada, showing stream sliding off top of esker. Photo by W. W. Shilts; used with his permission.

**Figure 4.** Sketch of small daughter esker diverging from and later rejoicing its parent esker. Based on photographs and field observations in Atnedalen, Norway. (From Hooke, 2005, Fig. 8.29).

**Flutes**

**Introduction**

Flutes are enigmatic geomorphic features of glacial landscapes. While they obviously reflect ice flow directions, certain aspects of their formation are less clear. There appear to be two basic varieties of flute: small and large. Small flutes were perhaps first described by Gilbert (1904). They are generally 0.2–2 m high and are frequently observed extending downstream from boulders lodged in the forefields of active glaciers (e.g. Benn, 1994). They tend to be uniformly tapered, and disappear after a few tens of meters.

Large flutes (the largest are called “megaflutes,” or “megalineations”) were apparently first noted by Tyrrell (1896). They typically have heights of a few to a few tens of meters, wavelengths (the transverse distance between ridge crests) of tens to hundreds of meters, and lengths of several kilometers. Bluemle et al. (1993) found that the stoss ends of many large flutes in North Dakota begin at thrust blocks which rise to the same height or slightly higher than the flute. The thrust blocks are chunks of Paleocene sandstone, a few tens of meters long and 2 to 4 m thick. Over much of their length, large flutes are not tapered. In Maine, we are concerned chiefly with these.

Isolated flutes are commonly flanked by troughs (Fig. 5) (e.g. Benn, 1995). It is generally inferred that the troughs are erosional, and that the material eroded from them was moved diagonally into or up onto the flute.
Figure 5. Schematic flow pattern of deforming till into cavity in lee of boulder. (Modified from Benn, 1994, Fig. 9).

**Origin**

One is tempted to infer that flutes are initiated by the obstructions (e.g. boulders or thrust blocks) often observed at their up-ice ends (Fig. 5), but such obstructions are not always apparent. Perhaps boulders can be removed by ice flow after flute formation, or perhaps the obstruction was a block of frozen till. Alternatively, perhaps such obstructions are not necessary for flute formation. This would be quite interesting.

Evidence for lateral flow of till into both small (Rose, 1989; Benn, 1994) and large (e.g. Jones, 1982) flutes has been provided by till fabric studies. The fabrics show a herringbone pattern, converging downflow. It has been inferred that the flow was a response to development of a cavity in the lee of an obstruction and was driven, in part, by shear imparted by the ice. High pore-water pressure weakened the till so it could flow.

Bluemle et al. (1993) found extensive evidence for such viscoplastic flow of till into a particularly long (27 km) large flute in North Dakota. Backhoe trenches across the flute revealed laminated silt and clay (lake deposits); gravelly sand, silt, and clay (till); and sandstone blocks derived from the bedrock. The lake sediments were commonly highly contorted. Some blocks, however, were not contorted; “intricate, essentially undisturbed bedding” was preserved in these. These blocks may have been frozen when emplaced.

Morris and Morland (1976) carried out an elegant theoretical analysis of flute formation. They found that the maximum stress driving flow of till into a cavity *increased* as the level of the till rose in the cavity. Thus, once stresses became high enough to initiate flow, a cavity of modest size would fill completely. In cavities more than a meter or so high, however, the greater density of the till (compared with the ice) prevented complete filling.

Why are large flutes long and uniform in height?

Fluidized till is denser than ice, so the stable geometry, if till were a Newtonian fluid, would be one in which the till formed a flat sheet under the ice. Thus, one might expect flow of till into a cavity to diminish as the ice, moving away from the initial obstruction, gradually closed the cavity.

Till, however, is a Bingham fluid with a finite yield stress. Morris and Morland (1976, p. 318) suggested that this would result in a limit, *on the order of a meter*, in the heights of flutes that could be formed by squeezing of till into a cavity. The stresses in higher flutes would exceed this yield stress, and the flute would collapse to the limiting height. Once at the limiting height, stresses would be below the yield strength and the till would “lock up,” forming a continuation of the initial obstruction. A cavity would continue to form downflow from the locked-up till. This
could lead to very long small flutes of quite uniform height. Such flutes are not observed; instead, small flutes are normally tapered.

A weakness of Morris and Morland’s model is that once the till comes into contact with the cavity roof they set the pressure on it equal to the ice overburden pressure. It seems likely, instead, that the bridging effect of a cavity’s arched roof will reduce the stress that the ice exerts on the till, thus allowing much higher flutes to develop.

Alternatively, Bluemle et al. (1993, p. 23) suggest that once in the cavity, till may lock up because the pore-water pressure in the lee of the obstruction is dissipated, increasing the till strength. Pore pressures might be reduced by dilatant hardening (Iverson et al., 1998), a process in which dilation increases the void space and thus decreases pore-water pressure.

**DEGLACIATION AND ICE-MARGINAL DEPOSITS**

**The coastal moraine belt**

Moraine ridges in Maine’s coastal lowland typically stand 1–10 m above the surrounding land surface, and are tens to hundreds of meters long. The smaller ones may be only 6–15 m wide, while some of the large stratified moraines found in eastern coastal Maine are more than 100 m across and can have lengths measured in kilometers. The Maine deglaciation chronology of Borns et al. (2004) suggests that the moraine belt was deposited between about 14,400 and 13,000 14C yr BP (~17.6–15.6 cal ka). These authors applied a reservoir correction of -600 years to the marine shell samples on which much of their coastal chronology was based. However, Thompson et al. (2011) obtained a correction of -1000 years based on dating of juxtaposed spruce logs and marine shells at the Mercy Hospital site in Portland. Application of this larger reservoir value, as proposed by Thompson et al. (in prep.), would bring the coastal Maine deglaciation chronology into closer alignment with the North American Varve Chronology elsewhere in New England (Ridge et al., 2014).

The moraines were formed along an ice margin in contact with shallow marine waters. Exposures in shoreline bluffs and gravel pits have shown marine sediments interbedded with the distal portions of morainal deposits, providing conclusive evidence that the ice margin did in fact stand in the sea. Small moraines are known by various names such as “washboard moraines”, “De Geer moraines”, or simply “minor moraines”. We refer to them as De Geer moraines. They are composed mostly of till and are common in mid-coastal to southwestern Maine, including the Searsport area visited during this trip. They presumably formed in the winter, when the ice front stabilized or readvanced slightly. This activity is recorded by various deformation structures seen in pit faces. In some areas, meltwater issued from beneath the retreating glacier under a high potential gradient, carrying quantities of gravel, sand, and mud and forming large stratified moraines. Lodgement till is commonly plastered against the stoss sides of these moraines (Thompson, 2015).

In clusters of De Geer moraines in the coastal zone, the spacing between moraine ridges has been reported to be 76 m in the southwest (Bloom, 1960); 66–93 m (70 m ave.) in midcoast Maine (Jong, 1980); and 50 m east of Penobscot Bay (Thompson, 1982). However, moraine spacing “varies considerably” in areas where they do not form distinct clusters (Smith, 1982). Analysis of LiDAR data enables more quantitative and accurate assessment of moraine spacing and other spatial properties (e.g. studies by Bouvier et al. (2015) in Sweden, and by Ojala et al. (2015) in Finland). The first analysis of De Geer moraine clusters shown on Maine LiDAR imagery was carried out by Eusden (2014). He documented moraine spacings ranging from 111 to 186 m and varying with slope aspect.

Hunter and Smith (2001) proposed that the “morainal banks” are annual features and thus provide a measure of deglaciation rates. They stated that the moraine spacing indicates retreat rates of 60–130 m/yr in southwestern coastal Maine, 80–130 m/yr in the midcoast region, and 50–125 m/yr in eastern Maine. We agree that the moraines are approximately annual and thus provide a measure of deglaciation rate. In the Penobscot Lowland, a rate of 150 m/yr would be broadly consistent with 14C dates (Hooke et al., in press). However, owing to the low resolution of techniques available for dating glaciomarine sediments, together with the likelihood of double moraines when retreat rates were low and missing moraines when they were high, the annual formation model is approximate.

Although the moraine ridges in southern Maine generally trend east to northeast, the marine-based tidewater glacier margin had a lobate shape. In some places, diachronous ice flow in neighboring ice lobes caused them to
overlap and produce cross-cutting moraines. Moraines and striations reveal that the glacier withdrew more quickly in certain coastal valleys. This probably occurred because of more rapid iceberg calving in the deeper water along the valley axes. The result was that ice flow converged into the valleys from either side. This process has been documented in the lower Penobscot River valley (e.g. Syverson and Thompson, 2008; Syverson and Olson, 2011), and in the lower Kennebec valley (Thompson, 2009a,b).

**Figure 6.** Schematic cross-section of ice-marginal glaciomarine deposits. From oldest to youngest, bedrock (bottom) is overlain by till (rocky pattern), sand and gravel (diagonal lines), delta topset gravel (circle pattern), and marine mud (Presumpscot Formation; dashed pattern). Diagram by W. B. Thompson.

**Fans and deltas**

Where the glacier margin stood in the ocean and remained stationary for a time, perhaps for a few years, coarse sediments accumulated rapidly at the mouths of ice tunnels. These deposits formed mounds of sand and gravel called subaqueous fans on the sea floor (Fig. 6). In places where the ice margin stabilized for longer periods, the fans built up to the ocean surface and became flat-topped glaciomarine deltas. Unlike the huge delta of the Mississippi and those of many other modern rivers, these glacial deltas consist mostly of coarse-grained sand and gravel. They have been called “Gilbert deltas” in reference to G. K. Gilbert, who described deltas of this type that were deposited into ancestral Lake Bonneville in Utah during the cool, wet Pleistocene climate.

A cross section through a typical marine delta reveals horizontal fluvial topset beds that overlie inclined foreset beds (Fig. 6). The topset beds were deposited in braided stream channels that shifted back and forth across the delta surfaces. These beds usually consist of coarse gravel. The foreset beds are composed of finer-grained sand and gravel that reached the ends of the channels and cascaded down the face of the delta. One can determine the approximate position of local relative sea level when the delta was built by measuring the elevation of the contact between the topset and foreset beds.

Thompson *et al.* (1989) identified, named, and described 101 glaciomarine deltas in Maine, and additional deltas have since been discovered through the MGS surficial geology mapping program. Of the deltas examined for the 1989 study, 69% are ice-contact deltas, 26% are “leeside deltas” (deposited in the lee of ridges protruding above sea level), and only 5% are valley outwash deposits built into the sea at some distance from the ice margin. Examples of the first two categories will be seen on this trip at Stops 1, 4, and 5.

**THE PENOBSCOT BAY PROBLEM**

In informal conversations, glacial geologists have pondered whether a calving bay developed in the ice sheet margin as the Laurentide Ice Sheet retreated across the Penobscot Lowland. Margin retreat by calving is generally considerably faster than by melting alone, and is faster in deeper water, so this is an eminently logical suggestion. Syverson and Thompson (2008) measured striations and small craig-and-tail features in the vicinity of Bangor and found good evidence for a remarkable late-glacial change in ice-flow direction from about N-S to roughly WNW-ESE within a couple of kilometers of the Penobscot River. The craig-and-tail features were interpreted to indicate westerly ice flow east of the river and easterly ice flow west of the river, suggesting a calving bay. The bay was inferred to be less than 2 km wide at that point.
Northeast of Bangor, however, the Katahdin esker veers to the southeast out of the Penobscot Lowland and across a range of hills (Fig. 7). According to our present understanding (e.g. Fig. 2a), this would only have occurred if the ice-surface slope were roughly parallel to the trend of the esker. This ice sheet topology had to exist throughout the time that the NW-SE trending part of the esker was deposited. As the esker was likely formed in segments 5–10 km long extending up-ice from the glacier margin, this slope must have persisted as the ice retreated across this part of the Lowland. This limits the extent of any calving bay at Bangor.

LiDAR imagery from either side of Penobscot Bay reveals many small De Geer moraines. Some of these are shown schematically in Figure 7. These moraines suggest that ice flow was to the S and SSE across Penobscot Bay, and do not provide support for a calving bay. If the SSE orientation becomes more southeasterly further north, it would, however, provide an explanation for the course of the Katahdin esker across the range of hills shown in Figure 7. Thus, it seems likely that any calving bays that developed were local and short-lived, and did not significantly affect the overall northward retreat of the ice across the Lowland.

**Figure 7.** Map of southern part of Penobscot Lowland. Light gray shading shows area below marine limit. Short (red) lines show approximate orientations of De Geer moraines.

**LATE-GLACIAL MARINE SHORELINES**

As relative sea level fell, every place below the upper marine limit was “shorefront property” at some point in time. Wave-washed sands and gravels at the marine limit are most likely beach deposits, but at lower elevations it is
often hard to determine whether such materials formed along the shoreline or slightly offshore in shallow water. So in many cases the latter deposits are shown on geologic maps as “marine nearshore deposits”.

A few paleoshorelines can be seen on air photos as linear zones where the bedrock has been stripped bare by wave attack. Each of these shorelines follows the contour of the land, having developed when local sea level was at a particular elevation. However, most are subtle features that are not readily apparent on air photos, so it is enlightening to see the many shorelines – both high-stand and regressive – that have been revealed by LiDAR imagery.

Figure 8 shows a striking LiDAR expression of moraines and a paleoshoreline near the east side of Penobscot Bay. A wave-cut scarp extends completely around Grey Ridge, outlining an island that existed during the high stand of sea level. This unusual image also reveals contrasting types of moraines above and below the marine limit. The moraines deposited in the sea exhibit a pattern and spacing that is typical of many other swarms of moraines in coastal Maine. However, the subaerial moraines on Grey Ridge show a dense pattern with short narrow ridges having less continuity and a slightly different orientation than those below the marine limit. Thompson (2011) illustrated a remarkably similar situation on Demuth Hill in Waldoboro. Past field studies have shown relatively few moraines above the marine limit in southern Maine, so further work is needed to understand the glacial processes responsible for the relationships seen here.

Figure 8. Lidar image of moraines (east-west ridges) and a paleoshoreline (upper-center) in Penobscot, Maine. The shoreline is at the upper limit of marine submergence and completely surrounds Grey Ridge. The area seen here is ~3.2 km across in E-W direction.

Shoreline deposits derived from till usually consist of angular, poorly sorted gravel that was not transported and abraded to any great extent. Those formed on submarine fans and deltas have better-rounded gravel and a higher percentage of sand inherited from their parent material. The latter type is apt to exhibit good shoreline morphology because it developed on material that could be eroded more quickly than till. Prominent wave-cut scarps can be seen on the seaward faces of some glacimarine deltas such as the Searsport delta (Stop 1).

ROAD LOG

Meeting point: Assemble at 8:30 AM in the parking lot at East Belfast School. If you are driving from the south on coastal Route 1, turn L on Route 141 (Swan Lake Avenue) just after crossing the bridge over the Passagassawakeag River in Belfast. The East Belfast School is located on your left, just 0.15 mile up the road (#14 Swan Lake Ave.). It takes at least 1.5 hours to reach Belfast from NEIGC headquarters in the Bath area, via U.S.
Route 1 and the Route 90 shortcut from Warren to Rockport. Come prepared with your lunch, water, etc. There will be one short walk in the woods; all other stops are easily accessed by car. The trip will progress westward and end around mid-afternoon. All stops are on private property, and permission must be obtained from the owners for any future visits!

**Mileage**  Note that cumulative mileages given here may differ from those shown on your odometer, due to variations in driving around pit areas, but the distances between stops are generally accurate.

0.0  Turn L out of school parking lot and go N on Rte. 141.
0.9  Turn R on Back Searsport Road.
2.6  Turn R on Skinner’s Pit Road. Drive past houses and park in gravel pit.

**STOP 1. SEARSPORT DELTA AND MARINE SHORELINE** (Skinner Pit, Belfast; Searsport Quadrangle).

![Figure 9](image-url) LiDAR image showing the locations of Stops 1-3 and some of the features discussed in the text. Triangle marks site of former pit exposure where the Searsport delta topset/foreset contact elevation was surveyed (Fig. 10). Area covered by this image is ~5 km across in E-W direction.

The Searsport delta is an ice-contact delta, having formed at the mouth of an ice tunnel. The path of the tunnel is now indicated by a chain of gravel pits in the esker that extends north from the northwest corner of the delta (Fig. 9). This esker is part of an esker system that has been traced discontinuously northward to the Newburg area, SW of Bangor. The LiDAR image in Figure 9 shows a prominent shoreline scarp at the marine limit. The scarp passes through Stop 2 and continues completely around the seaward edge of the delta and the higher till area to the north, outlining what was an island just after glacial retreat from this area. Figure 9 also shows a small area of regressive shorelines to the east of the pit at Stop 1.
Pits excavated in this delta in the past are only 3-6 m deep (Thompson and Locke, 2013), suggesting that the deltaic sand and gravel is thin in comparison to other Maine deltas of similar surface extent. The depth of past workings, however, may have been limited by the water table, as some recent sand and gravel extraction has occurred from ponds in the pit complex. The till seen on the hillside just north of the delta is a lodgement deposit and probably extends southward beneath the delta at a shallow angle, in which case the downward movement of ground water would be restricted by the low permeability of the till.

The pit at this stop is located in the proximal part of the Searsport delta. A bedrock outcrop was formerly exposed next to the pit access road, where there is now a small pond. Striations on the ledge surface indicated glacial ice flow toward the SE (143°). The present pit face shows very coarse, poorly sorted pebble-boulder gravel with weak sub-horizontal bedding. Boulders up to at least 1 m are present. This gravel is interpreted to be part of the topset (glaciofluvial) portion of the delta. A deeper exposure at the south end of the pit shows dipping planar-bedded sand and gravel interpreted as foreset beds. Thompson (2014c) examined other workings SW of here, in the distal part of the delta, where the material is finer grained and relatively well sorted and stratified. Many exposures in the latter area show either foreset beds or (near the seaward rim of the delta) sand and gravel interpreted to be a shoreline deposit resulting from reworking of deltaic sediments.

The delta study by Thompson et al. (1989) includes a surveyed elevation of 89.3 m (293 ft) for what was thought to be the contact between topset and foreset beds in the Searsport delta. In 1993 Thompson re-examined the part of the delta south of today’s stop and found a prominent meltwater channel on what was then a blueberry field on the delta plain. The channel was 2 m deep, “hundreds of feet long” (WBT field notes), and trended SSE across the delta to the point where it was intersected by a fresh pit face. The section through the axis of the channel floor exposed ~ 1 m of coarse topset gravel overlying 30 cm of silt-sand-pebble gravel interpreted as a slack-water deposit. The silt resembled fine-grained deposits found along topset/foreset contacts in other deltas, and may have been a tidal facies. A second pit nearby revealed a topset/foreset contact, with both units being coarse gravel (Fig. 10). A precise survey of this contact yielded an elevation of 91.0 m (298.4 ft), which is thought to be the best record of late-glacial relative sea level in the Searsport area. The survey site is shown on the surficial geology map by Thompson (2014c).

Figure 10. Pit face exposing subtle contact between topset (top) and foreset (bottom) units in the Searsport delta. The line marks the contact. Photo taken in 1993 by W. B. Thompson.

In past years, gravel pits have also revealed excellent exposures of shoreline features on the edge of the Searsport delta. Figure 11 shows marine gravel deposited on a wave-cut platform along the seaward delta margin.
Figure 11. Pit face on SE corner of Searsport delta, showing marine shoreline or nearshore gravel unconformably overlying eroded delta foreset/bottomset sand beds with large-scale water escape structures. Photo taken in 1984 by W. B. Thompson.

STOP 2. FLUTED TILL SURFACE AND MARINE SHORELINE (Searsport; Searsport Quadrangle).

The clearing in the cemetery property shows a good example of a long, low flute. The flute extends 1.5 km NNW from here (Fig. 9). The raised marine shoreline seen on the LiDAR image (Fig. 9) crosses the entrance to the cemetery road. Here the shoreline is subtle and could easily be overlooked if LiDAR were not available. Rounded stones are present where a small brook intersects the cemetery road on what appears to be the shoreline terrace. Based on observations along other raised marine shorelines in coastal Maine, coarse gravel is likely to occur in places like this, where till has been eroded by wave action.

STOP 3. SEARSPORT MORAINES (Searsport; Searsport Quadrangle).

This brief stop will give you an opportunity to see one of the De Geer moraines. Depending on the number of cars, we may first drive across a couple of the moraines before we turn back and walk out on one. It is easy to see how moraines like this were missed by earlier mappers lacking LiDAR images. De Geer moraines in Maine are commonly asymmetric. Submarine examples like this one tend to be steeper on their proximal sides, likely owing to ice push on the proximal side and submarine-fan type deposition on the lee side. Subaerial De Geer moraines tend to be steeper on their distal sides owing to an ice-dozer effect forcing material to tumble down a distal angle-of-repose slope.
6.7 Turn R onto U. S. Route 1 and go W, back to Route 141 in East Belfast.
10.7 Turn R onto Route 141 north.
17.0 Swanville settlement. Stay on Route 141, driving along W side of Swan Lake.
21.7 Take sharp L turn onto Back Books Road (no sign!).
21.9 Continue **straight ahead** at jct., going onto Dickey Hill Road (gravel road; no sign)
22.1 Turn L on Curtis Road (no sign).
22.7 Turn R onto gravel access road with sign for Michael Tripp Construction / Thompson Pit.
22.9 Park in Thompson Pit.

**STOP 4. IRISH HILL DELTA (Thompson Pit, Monroe; Brooks East Quadrangle).**

![LiDAR image](image_url)

**Figure 12.** LiDAR image showing locations of Stops 4 and 5 on the Irish Hill delta in Monroe. The area shown here is ~3.2 km across in E-W direction.

This gravel pit is located in the distal part of the Irish Hill delta (Fig. 12). The delta wraps around higher areas of till and bedrock that were islands during the high stand of sea level. It is a good example of a leeside delta, having formed where glacial meltwater issued from the gap between Irish Hill and Clement Hill (Thompson et al., 1989). In this case the feeder stream was subglacial, as indicated by a discontinuous esker ridge that passes through the gap and ends near the proximal margin of the delta. There must have been at least a tongue of ice extending seaward between the hills when the esker formed, but the hills themselves reach elevations of 649–810 ft (well above the marine limit) and would have been at least partly ice-free by the time the delta was deposited.

There is a prominent and unexpected kettle between the esker terminus and the delta. Does this kettle simply mark a chance persistence of dead ice between the larger hills to the NW and the paleo island to the SE? Or did the kettle form by erosion at the site of a submerged jet (ice-marginal glacial “fountain”), the velocity of which was high owing to a high potential gradient at the end of the ice tunnel?
The Thompson pit exposes ~10–15 m of section in the delta topset and foreset beds. The topsets consist of up to ~2 m of massive, poorly sorted, pebble to boulder gravel. The underlying foresets are planar-bedded sand to cobble gravel and dip up to 30° toward the E to ESE. A recent fresh face in the SE corner of the pit showed a lateral truncation of the coarse topset gravel by finer-grained sand to pebble gravel in the seaward direction. The latter unit may be a reworked deposit formed by marine shoreline processes. Thompson et al. (1989) surveyed the topset/foreset contact in this pit and obtained an elevation of 96.0 m (315 ft) for local relative sea level when the delta was deposited.

23.1 Exit pit road and turn L on Curtis Road.
23.7 Take sharp L turn onto Dickey Hill Road (no sign).
24.3 Turn L into gravel pit (across from Mt. Solitude Cemetery)

STOP 5. JAMES TRIPP PIT (Monroe; Brooks East Quadrangle).

This is an optional stop that we may visit if time permits. It is a small gravel pit located in a slightly more proximal part of the Irish Hill delta relative to Stop 4 (Fig. 12). The excavation has exposed up to about 6 m of coarse gravel including boulders to 1 m in diameter. Many of the boulders are subangular to angular, so probably have not been transported far. The topset/foreset contact is indistinct, but in 2013 the upper SE side of the pit showed ~2.4 m of reasonably definite topsets consisting of massive, poorly sorted pebble-boulder gravel and minor sand. The SSW side exposed similar coarse gravel, but with faint bedding and interbedded pebbly sand. The latter material was thought to be part of the foreset unit.

24.6 Exit Tripp Pit and turn R on Dickey Hill Road.
25.3 Jct. with Back Books Road. Continue straight ahead, returning to Route 141.
25.6 Turn L on Route 141 and go north.
27 Watch for Doak Farm on R. Over the next 0.3 mi. several NNE-trending moraines can be seen in fields on both sides of Route 141. They are prominent on LiDAR but do not appear in the topo map contour pattern!
27.6 Monroe village. Turn L on Route 139.
31.2 Park carefully on R shoulder of road.

STOP 6. BASIN POND ESKER NET (Monroe; Brooks East Quadrangle).

Starting from Route 139, we will follow a woods trail NW along the crest of one of the esker ridges and observe its connections to other ridges in an anastomosing esker net (Fig. 13). The individual ridges in the vicinity of this stop are generally narrow and less than 10–15 m high. They are concealed by forest cover and are not evident on topographic maps.

Figure 13. Close-up of SE part of Fig. 14, showing ridges comprising the esker net (center to lower right) at Stop 6. Route 139 crosses the esker system in the lower right part of image.
Smith and Thompson (1986) showed a single esker in this area during reconnaissance geologic mapping of the Brooks 15-minute quadrangle. They may not have recognized the complexity of the esker net, but in any case they were limited in what could be depicted at the small scale of the published map.

Continue SW on Route 139.

31.5 Turn R on Pattee Road (gravel; sign faces W, so is not readable until you turn!).
32.6 Turn R on Pattee Road Ext.
32.9 Turn around at wide driveway entrance on L and park carefully on R (N) shoulder of road. A few cars can park on sides of driveway entrance area, but do not block this driveway!

**STOP 7. HASKELL HILL MELT WATER SCOUR ZONE (Monroe; Brooks East Quadrangle).**

Hillshade lidar images of the Brooks E-W and Unity quadrangles show two principal types of landscape above the marine limit: irregular areas of rough bedrock-controlled topography alternate with glacially smoothed, fluted till surfaces. The metamorphic bedrock in the N and W parts of the area is characterized by prominent NE-trending strike ridges oriented transverse to the SE flow of the Laurentide Ice Sheet during the last glacial maximum. Lodgement till was plastered against the proximal sides of these ridges, reaching thicknesses locally exceeding 40 m and producing smooth and often fluted surfaces. Meltwater channels cut into upland till surfaces, as well as scattered moraines, glaciomarine deltas, and subaqueous fans below the marine limit, collectively record the recession of the Laurentide Ice Sheet margin from the area.

Zones of erosion by subglacial meltwater – referred to here as scour zones – were revealed by the LiDAR imagery during surficial geologic mapping of the Brooks East and West quadrangles for the Maine Geological Survey (Thompson, 2014 a,b; Thompson and Hooke, 2014). One of the best examples occurs in the vicinity of this stop. The meltwater scour zones are generally parallel to regional S to SE-trending esker systems. On LiDAR imagery they appear as linear, often sharply-bounded, areas where most of the till cover has been eroded away. This process resulted in distinctive ribbed topography transverse to the glacier flow direction and reflecting the NE-trending structural grain of the underlying bedrock. The ribbed topography is especially noticeable where subglacial streams crossed bedrock strike ridges and incised the smooth till slopes that mantle the proximal sides of the ridges. The areas from which till was stripped are larger than typical subaerial meltwater channels and probably reflect a distributed subglacial stream system consisting of anastomosing broad low water courses (Fig. 1).

The scour zones may be as much as several kilometers long and 200 to at least 1400 meters wide. Their points of origin are not readily apparent on the LiDAR imagery. They typically begin in a diffuse area of bedrock-controlled topography encompassing much of the N and NW parts of the study area. This is consistent with a distributed, arborescent stream system with individual channels being braided and with discharges increasing downglacier. A thin patchy till cover usually remains over much of the scour zones.

In places the scour zones converge on rock-floored saddles (gaps) on bedrock ridges, consistent with Shreve’s (1972) model of the pattern of subglacial hydraulic potential contours described above (Fig. 2a). At the present stop on Pattee Road Extension, there are several channels cut into bedrock where glacial meltwater crossing the local saddle abruptly plunged downslope to the SE. Meltwater abrasion features occur on rock surfaces associated with the channels at this stop. A small narrow chute – seen just SE of the road – has smoothed undercut walls. Quartz veins are common in the bedrock here, and the knobby surfaces of some of these veins have likewise been worn smooth by sediment-laden streams.

As described above (Fig. 2), scour on the proximal sides and crests of bedrock saddles is predicted by the increase in hydraulic potential gradient resulting from ice flow against adverse slopes. Eskers commonly occur on the distal sides of such saddles, in some cases forming anastomosing networks (esker nets) that are characteristic of parts of esker systems formed near a glacier margin. A good example of one of these esker nets was seen at the previous stop.
Figure 14. Lidar image of Haskell Hill meltwater scour zone (upper left) and esker net (lower right) in NW part of Brooks East quad. Stop 7 is located in center of image.

The processes and length of time involved in formation of the scour zones, and their relationship to esker deposition, are poorly understood. Our rough calculations suggest that many may have formed in a couple of decades. Subglacial channels cut into bedrock are rare. This is probably due to the high sediment loads common in
subglacial streams (due to melting of dirt-bearing ice in conduit walls), together with frequent shifting of stream courses. The high sediment loads result in deposition (eskers) rather than erosion.

**End of trip.** If heading southbound on I-95, you can drive west from our last stop on Route 139, go 28 miles/40 minutes, and pick up the Interstate in Fairfield.

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