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T. K. Weddle

Michael Retelle
Bates College, mretelle@bates.edu

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GLACIOMARINE DEPOSITS OF THE LATE WISCONSINAN CASCO BAY SUBLOBE OF THE LAURENTIDE ICE SHEET

by

Thomas K. Weddle, Maine Geological Survey, State House Station 22, Augusta, ME 04333
Michael J. Retelle, Department of Geology, Bates College, Lewiston, ME 04240

INTRODUCTION

Retreat of the Late Wisconsinan Casco Bay Sublobe of the Laurentide Ice Sheet in southwestern Maine is represented by deglacial-phase deposits which include glaciomarine deltas, fans, stratified end moraines, and fossiliferous glaciomarine mud, as well as nearshore deposits associated with synglacial sea level and postglacial uplift. This trip will examine exposures illustrating the style and timing of ice-marginal recession through the lower Androscoggin River valley about 14,000 to 13,000 yr BP, and subsequent regression of the sea which occurred as late as 11,600 yr B.P. If exposures permit, we will visit and discuss the significance of the extensive Brunswick sand plain, which we interpret as a coastal braid delta (Blair and McPherson, 1994). This landform is comprised of fluvial deposits over nearshore regressive sediments. Radiocarbon age-dates on nearby regressive deposits infer that the delta built to a regressive but periodically stable sea level sometime after 12,800 yr BP, probably during the time of meltwater pulse IA (Fairbanks, 1989), and that most of the region above 36 m asl was emergent by that time.

Timing of late-glacial events and ice retreat in the region has been better constrained by datable materials found in association with ice-marginal positions (Weddle and others, 1993, 1994; Weddle, 1994). The precisely obtained altitudes of ice-marginal deltas deposited during systematic ice retreat provide a record of sea level fluctuations during this time, and characterize the nature of postglacial uplift in the region (Thompson and others, 1989, Koteff and others, 1993; cf., Barnhardt and others, 1995). The following section describes our understanding of deglaciation in southwestern Maine to date.

LATE WISCONSINAN GLACIAL GEOLOGIC HISTORY

The Laurentide Ice Sheet reached its maximum position 300 to 400 km southeast of the present Maine coast on the continental shelf at least by 26,000 yr B.P. (King and Fader, 1986). The ice mass in New England is subdivided into various lobes (Stone and Borns, 1985). The lobes in the western Gulf of Maine include the Cape Cod Bay lobe and the South Channel lobe. From this terminal position, the lobes retreated northward, crossing the present southwestern Maine coast by 14,820 ± 120 yr B.P. (AA-10166; Weddle and others, 1994). By this time, smaller ice sublobes filled the valleys, and in the field trip area, we propose the name Casco Bay sublobe for the ice mass which occupied the lowland region around Casco Bay, including the modern drainage areas of the lower Androscoggin River and the Royal River.

The northward retreat of the ice was accompanied by a marine transgression across the isostatically depressed crust, and is marked by numerous stillstands represented by end moraines, grounding-line fans and deltas, and at the inland marine limit, large, ice-contact glaciomarine deltas and broad, outwash plains. In coastal Maine, fossiliferous marine mud is interbedded with subaqueous outwash and diamicton, composing the large end moraine complexes. Invertebrate fossils inside the moraines yield a clear picture of temporal and spatial conditions at the grounding line. In addition to chronologic control, the faunal assemblages of both macro- and microfossils allow paleoceanographic reconstructions to be made (Dorion, 1993, 1994; Kaplan, 1994; Kreutz, 1994; Dorion and others, 1995). The radiocarbon age-dates are from fossil assemblages that define the timing of the northward retreat of the ice (uncorrected for ^{14}C-marine reservoir effect; cf., Mangerud and Gullickson, 1975). The age date of 14,820 ± 105 yr B.P. is from Scarborough in southwestern Maine, and is the oldest radiocarbon age-date on marine fauna in the state. It matches numerous other radiocarbon age-dated fossils in western and eastern coastal areas. Kelley and others (1992) obtained an age of 14,090 ± 450 yr B.P. (PITT-0743) on *Macoma balthica* from glaciomarine sediment now submerged by Holocene sea level rise. The core location is off Prouts Neck in 21 m of water depth. Radiocarbon age-dates from *Portlandia arctica* fossils onshore in ice-proximal deposits ranging from about 14,000 to 12,900 yr B.P. from Freeport to Lewiston, respectively, delineate the timing of grounding line positions in that area (Weddle and others, 1993; Weddle, 1994). In addition, because the moraines trend east-west, ice retreat was nearly due north. Radiocarbon age-dates to the south should be older, and conversely, age dates to the north should be younger, with the *caveat* that the stratigraphic settings in each case must be consistent. The moraines in eastern Maine share the same east-west orientation as western coastal Maine (Kaplan, 1994). Together, it is then inferred that ice retreat in both western and eastern coastal Maine was generally northward in sequential style. Additional radiocarbon age-dates from the present Maine coast northward to the inland marine limit show relatively faster retreat in large marine re-entrants in the Penobscot and Passamaquoddy embayments (Dorion, 1994; Dorion and others, 1995). This style of retreat is seen on a smaller scale in eastern Maine, where in topographically low valleys, retreat is increased relative to adjacent uplands where the ice margin remained pinned for a period of time (Kaplan, 1994). However, Lowell and Borns (1994) have suggested that the effectiveness of a calving embayment in the Penobscot River valley was impeded by topography and shallow water depths of the transgressing ocean, and that deglaciation may have been more influenced by climate.

The transgressing sea reached the inland marine limit in the field trip area (85 - 95 m asl) probably between 14,000 and 13,500 yr B.P. The marine transgression ended as isostatic uplift exceeded sea level rise. An AMS-radiocarbon age-date of 13,300 ± 50 yr B.P ($\delta^{13}C$ = -0.96 ppt, OS-4419) on *Mytilus edulis* from a nearshore deposit at 60 m asl in Pownal infers regression in the region was underway by that time. The shoreline regressed to near 30 m asl in the lower Androscoggin River valley by approximately 12,800 yr B.P. (12,820 ± 120 yr B.P. (SI-7017), Retelle and Bither, 1989; 12,850 ± 45 yr B.P. (OS-2348), Weddle, 1994), and emergence was near complete by as late as 11,500 yrs B.P. (Smith, 1985; Smith and Hunter, 1989; Anderson and others, 1990).

Deglaciation Style

The nature of retreat of the late Wisconsinan ice sheet from terrestrial southern New England has been summarized by Koteff and Pessl (1981). They characterized the Laurentide ice margin as active ice, systematically retreating northward with a narrow stagnant ice margin flanking live ice, and termed this mode of deglaciation "stagnation-zone retreat." As the margin retreated, deposits known as morphosequences were laid down by meltwater, and they record a relative chronology of ice recession and deposition (Koteff, 1974). The width of the stagnant zone varied depending on local topography, and generally appears to have been only a few kilometers. Minor ice readvances have been found, but the pattern of systematic and steady ice retreat is persistent throughout the region.

The morphosequence concept is equally useful in the glaciomarine environment, especially in coastal Maine where the marine-based ice sheet was grounded rather than floating. Although a marginal stagnation zone may not have existed along the marine-based margin, features such as the numerous washboard moraines and ice-shoved proximal portions of deltas and fans attest to nearby active ice throughout the deglacial history of the area. The glaciomarine fans and deltas associated with the ice margin were deposited from subglacial tunnels or from glacial streams at their termini. Ice-marginal deglacial-phase glaciomarine deposits in coastal Maine, have been described in detail by Stemen (1979), Jong (1980), Ackert (1982), Lepage (1982), Smith (1982), Smith, Stemen, and Jong (1982), Thompson (1982), Miller (1986), Smith and Hunter (1989), Retelle and Bither (1989), Thompson and others (1989), and Ashley and others (1991). The ice-marginal deposits to be seen during this field trip are similar in their internal sedimentological character to the deposits described in the cited studies as examples of deposits from a temperate marine grounded ice-sheet. An excellent summary of the glacial geology of southwestern Maine can be found in Thompson and others (1995), which is reprinted, in part, in this guidebook as trip A-4.

Morphostratigraphy. The systematic northward retreat of the marine ice sheet is documented by a complex association of sediments and landforms deposited along the former grounding lines of the tidewater glacier margin throughout the coastal lowland. The most common ice-marginal deposits are composites of linear end moraines, submarine fans, and deltas, which are products of meltwater discharge into the marine basin at the grounding line.

End moraines. There are thousands of end moraines in the coastal lowland and they occur in clusters of parallel ridges, many concealed beneath younger ice-marginal deposits or blanketed by fine-grained glaciomarine mud (Smith, 1982; Thompson, 1982). The principal characteristic shared by most moraines is their ridge morphology, or at least a linear alignment along a former ice-marginal position. In general, the moraines are 3 - 45 feet (1 - 15 m) high, and 15 - 45 feet (5 - 15 m) wide, with some larger moraines exceeding 300 feet (100 m) in width. Most in southwestern Maine are minor moraines, referred to as washboard or DeGeer moraines (Smith, 1982; 1985) and are typically a few feet high to several hundred feet long and are regularly spaced around 200 - 250 feet (65 - 70 m). Larger moraines are generally more complex in morphology and composition. Many of the moraines are comprised of diamicton including lodgement till, debris flow deposits and glaciotectonized stratified drift. The larger landforms, commonly referred to as stratified end moraines (Borns, 1973; Ashley and others, 1991) consist of stratified sand and gravel interbedded with minor amounts of diamicton. The stratified moraines usually exhibit evidence of glaciotectonism, with the interstratified materials folded and thrust-faulted by overriding ice or ice-shove. The occurrence of the many moraines and the intimate stratigraphic relationship with glaciomarine sediments indicate that: (1) the grounded ice margin was active during its northward retreat, and (2) ice retreat was accompanied immediately by marine incursion of the coastal lowland as ice was in contact with the sea in a calving bay regime (Smith, 1982; Hughes and others, 1985).

Eskers, submarine fans, and deltas. Sediment deposited by meltwater occurs in several genetically connected and related landforms in the ice-marginal zone and include eskers, submarine fans, and deltas. Eskers (Thompson and Borns, 1985a) ice-tunnel deposits (Ashley and others, 1991) or subglacial conduit facies (Sharpe, 1988) are terms used to describe the coarse-grained sediments from a predominantly high-energy fluvial source that delivers sediments to the ice-proximal zones of fans and deltas. These deposits occur as distinct and separate sinuous ridges in valleys above marine limit (cf. Thompson and Borns, 1985a), as feeder "tails" on ice proximal sides of deltas (Thompson and others, 1989), and as coarse-grained cores of fans and deltas where the supporting ice has retreated and a delta or fan landform has prograded basinward over its former conduit. Sollid and Carlson (1984) describe eskers and DeGeer moraines from Norway, which may be analogous to the glaciomarine deposits associated with the "beaded" esker systems of Maine.

Submarine fans originate at the mouth of the esker or meltwater conduit and grade distally to the seafloor. The submarine fans, also referred to as submarine outwash (Rust and Romanelli, 1975), subwash fans (Burbidge and Rust, 1988), and grounding line fans (Powell, 1990) are irregular mound, cone, or fan-shaped features that commonly drape or flank the moraine ridges. Over time, particularly if the retreat of the grounded ice margin is halted at a pinning point on a subglacial topographic high or a valley constriction, a submarine fan may aggrade vertically and prograde distally and evolve into a more massive and extensive deposit that approaches and may eventually even reach contemporaneous sea level. In this latter and most developed case, the fan may eventually become a delta with a flat subaerially exposed topset plain (Powell, 1990; Slayton, 1993). Isotopic analyses on marine fossils from eastern Maine by Kreutz (1994) infer that paleoceanographic conditions support an ice-marginal model similar to that of Powell (1990), where meltwater emanating from the base of the ice sheet rises directly to the surface as a plume.

The fan sediments represent the transition between fluvially dominated processes associated with the ice tunnel environment, and processes of the proglacial marine basin. Consequently, materials in the fans show complex stratigraphic relationships both parallel to the ice margin and distally from the former tunnel mouth. Sediments in the proximal zone include coarse-grained gravelly stratified materials supplied by the meltwater currents and diamicton that may originate from slope failure and downslope movement from the adjacent moraine, or ice thrust (Retelle and Bither, 1989; Ashley and others, 1991). In medial and distal zones of the fans the major facies include rhythmically bedded sand and mud that grade to the muddy seafloor. These sediments, termed cylopsams and cyclopels (Mackiewicz and others, 1984) were likely deposited by suspension from a highly turbid but buoyant suspended sediment (Cowan and Powell, 1990). Resedimentation of these deposits, particularly in the proximal and medial portions of the fan produces grain flows and debris flows of varying textural composition. Stratigraphic sections in the medial and distal portions of the fans are commonly fining-upward sequences (Retelle and Bither, 1989) that grade transitionally upwards from coarse proximal fan sediments at the base to distal fan sediments. The fining-upward sequence may either represent ice retreat with removal of the ice tunnel sediment source or a lateral switching of a distribution channel on the fan lobe.

Over 100 glaciomarine deltas occur in southern Maine. The origin and stratigraphy of these deltas has been described and categorized by Thompson and others (1989). Although they are widely distributed across the coastal lowland, the largest examples are located at or near the inland marine limit in southwestern and eastern Maine. Many are localized where the ice margin was temporarily pinned against bedrock ridges and other topographic highs. Many pit exposures in deltas show that they are Gilbert-type deltas, with horizontal fluvial topset beds (delta-plain deposits) overlying inclined foreset beds deposited on the prograding delta front. These deltas are typical of environments where coarse-grained sediments are rapidly deposited in basins of sufficient depth to produce a steeply-sloping delta front (Nemec, 1990).

The contact elevation between the topset and foreset beds in Maine's deltas approximates the late-glacial sea level to which the deltas were graded. Thompson and others (1989) surveyed the topset/foreset contact elevations to define the plane of the upper marine limit. Isostatic uplift has tilted the plane to the southeast, and it now has an average slope of 2.82 ft/mi (0.53 m/km) in the Kennebec River Valley region. However, this is only a minimum gradient for postglacial tilt, because any uplift during ice retreat would have lowered relative sea level and reduced the slope of the delta elevation profile. An estimate of postglacial tilt in the lower Androscoggin River Valley using elevations of delta tops from topographic maps and topset/foreset contact measurements from Thompson and others (1989) is between 3.33 and 3.75 ft/mi (0.63 - 0.71 m/km). Koteff and others (1993) have determined the postglacial tilt from deltas in coastal New Hampshire and southwestern Maine to be 4.5 ft/mi (0.85 m/km), supporting their proposal for delayed uplift in southern New England (Koteff and Larsen, 1989). The discrepancies between the regions are partly the result of recent topset/foreset contact measurements and reinterpretation of previously measured delta elevations in the Maine - New Hampshire border area (Koteff and others, 1993). Moreover, if uplift had begun by the time the ice margin was in the lower Androscoggin River Valley, a less steep postglacial gradient of the water plane would be expected in the field trip area.

Recent work has shown that the deltas in extreme southern Maine were inundated by the rising late-glacial sea (Koteff and others, 1993; Weddle and others, 1993). The deltas have been wave eroded to varying degrees, including partial or even complete removal of the topset beds. Consequently, shoreline or nearshore deposits of sand and gravel may unconformably overlie the eroded delta foresets, producing a stratigraphic relationship which may at first be confused with topset/foreset contacts. This reworking by marine processes also explains the scarcity of preserved meltwater channels on delta tops in southwestern Maine. Glaciomarine deltas elsewhere in the state have distributary channels on the delta plain, and local terracing of channels on some of the larger and longer-lived deltas in eastern Maine (e.g., Silsby Plain and Pineo Ridge) indicates that uplift exceeded eustatic sea-level rise as the deltas were forming.

Distal outwash deltas, formed where meltwater streams entered the sea some distance from the ice margin, are rare in Maine. Isostatic uplift was in progress across much of the region as the ice withdrew inland from the marine limit. Any outwash reaching the sea was distributed as a thin fluvial sand across the glaciomarine mud of the emergent sea floor as relative sea level dropped in response to uplift. An example of this type of deposit is the Brunswick sand plain (Weddle, 1994; Androscoggin sand plain of Leavitt and Perkins, 1935). The plain surface grades from about 36 m to 18 m asl (surface gradient 0.028). Shallow excavations in the surface of the plain reveal fluvial trough-cross beds of fine- to medium-grained sand, in places containing mud rip-up clasts and mud drapes, typical of a braided-stream environment. The plain morphology and sedimentology classify it as a coastal braid delta (McPherson and others, 1988; Blair and McPherson, 1994), formed as braided streams from the late- or post-glacial Androscoggin River valley entered the regressive sea.

These braided-stream deposits unconformably overlie a coarsening-up sequence of massive and laminated mud grading upwards to interbedded silt and fine- to medium-grained sand layers, known collectively as the Presumpscot Formation (Bloom, 1960,1963). This muddy unit beneath the fluvial sand represents a transition during glaciomarine transgression from a distal glaciomarine and submarine plain environment to shallowing conditions during marine regression. Subsequent to the deposition of the massive glaciomarine mud, existing units were reworked by the marine regression and nearshore deposits were laid down. In the study area, these deposits are best exposed at eroded coastal bluffs (Weddle and others, 1993; Weddle, 1994), and are reported from detailed geotechnical logs from the Superfund site at the Naval Air Station in Brunswick (Draft Final Feasibility Study, ABB Environmental Services Inc., November, 1991, Portland, Maine).

Several origins for the Brunswick sand plain may be proposed: increased discharge by the late- to post glacial Androscoggin River, deposition due to loss of capacity where the late- to post-glacial Androscoggin River exited from a confined valley to an unconfined valley, and in association with the above, the plain was deposited during a period when falling relative sea-level may have stabilized as a response to a rising eustatic sea-level, balancing glacio-isotatic uplift long enough for the coastal braid delta to form. In any model, the age of the plain is minimally constrained by the elevations of radiocarbon-age dates on *Hiatella arctica* shells (13,100 ± 125 yr B.P., $\delta^{13}C = 1.1$ ppt, GX-20774) found beneath the outer edge of the plain (pers. comm., A. M. Hussey, II, 1995), and from sites beneath regressive deposits in nearby Harpswell (30 m asl) and Topsham (47 m asl), which constrain the age of the plain to be as old as or slightly younger than 12,500 yr B.P. (Retelle and Bither, 1989; Weddle, 1994).

Other broad sand plains are present in Maine, although they are at higher elevations than the Brunswick sand plain, and many are found at the inland extent of the marine limit, sometimes associated with ice-marginal deposits (upper elevation range only, e.g., Lexington Flats, 136 m asl; Oxford, 104 m asl; Gray, 98 m asl; Lyman - Limington, 91 m asl; Silsby, 88 m asl; Turner, 82 m asl; Norridgewock 76 m asl; Deblois, 76 m asl; Berwick, 73 m asl; Sanford - Kennebunk, 72 m asl). The Brunswick plain is significantly lower, at about 36 m asl, and is not associated with ice-marginal deposits.

It is likely that the marine regression was influenced by eustatic sea-level rise, marked by periods of rapid increase (Fairbanks, 1989), and that a steady relative sea-level fall was not consistent during regression. In southwestern Maine, nearshore deposits derived from wave activity during marine regression are found at elevations near 30 m asl, particularly on drumlins (Leavitt and Perkins, 1935), although higher elevation nearshore deposits present on other landforms also are found (Weddle and others, 1993). Where detailed surficial geologic mapping has been done, morphologic evidence for stillstands that apparently occurred during the Late Wisconsinan marine regression in Maine is noted by Smith (Portsmouth 7.5' quadrangle, 1988, and in press), Hildreth (Biddeford 7.5' quadrangle, 1990), and by Clinch and Thompson (Prout's Neck 7.5' quadrangle, 1990). Prominent deposits or erosional features noted by Smith (1988) and Hildreth (1990) are found at a range of elevations from 66 m asl to 6 m asl.

In the town of Scarborough along the Nonesuch River valley, approximately 50 km southwest of Brunswick, Clinch and Thompson (1990) report a thick blanket of fluvial sand that is underlain almost solely by the Presumpscot Formation. The modern Nonesuch River is underfit with respect to the fluvial deposits in its valley, and Clinch and Thompson (1990) postulate that the sand may represent an early postglacial channel for the Saco River, beheaded by headward erosion along the modern lower course of the Saco River. Farrell (1972) describes from cores and natural exposures in the lower Nonesuch River valley and Scarborough Marsh fluvially deposited sand underlying the marsh peat and overlying the Presumpscot Formation, laid down when the sea stood 12 to 15 m above present sea level.

In the adjacent Old Orchard Beach quadrangle to the west, the same fluvial deposits in the Nonesuch River valley can be traced to elevations of 36 m asl (Retelle, 1991), the same elevation as the upper range for the Brunswick sand plain. Also in that quadrangle, at a fossil-mammoth locality (Caldwell, 1992) adjacent to the Nonesuch River and at about 30 m asl, a detailed stratigraphic and paleoenvironmental study of the fossil site records a gradational upsection sequence of *transgressive* glaciomarine mud to *regressive* nearshore deposits, stratigraphically overlain by fluvial deposits of the Nonesuch system (C. C. Dorion, pers. comm., 1994).

The mammoth remains were recovered from the top meter of the *regressive* sequence of marine rhythmic laminites of sand and mud. An AMS radiocarbon age-date on *Portlandia arctica* from the *transgressive* mud below where the mammoth remains were found is reported by Weddle and others (1994) at 14,820 ± 105 yr B.P. (AA10166). Radiocarbon age-dates from nearby deposits that probably represent regressive assemblages to the north and east of the site reported by Stuiver and Borns (1975) and Smith (1985) range in elevation from near 40 m asl to 18 m asl, and in age from approximately 12,400 to 11,400 yr B.P. Downstream from the mammoth site, Clinch and Thompson (1990) map several terraces that were formed along the Nonesuch River valley, as well as Pleistocene shoreline deposits in the quadrangle, which demonstrate that there were marked interruptions in the marine regression.

Presumpscot Formation. Glaciomarine mud has been named the Presumpscot Formation in Maine by Bloom (1960; 1963); the unit has been mapped and extended into New Hampshire by Koteff (1991). Subsurface data and surface exposures show that this unit directly overlies bedrock, till, fans, and end moraines, and is interbedded with medial and distal components of fans and deltas. It can be massive or layered, containing outsized clasts and ice-rafted debris, and is fossiliferous, with a rich assemblage ranging from foraminifers and diatoms through mollusc shells to marine vertebrates. Based on associated fossil assemblages, it is considered a late Pleistocene cold-water *marine* unit (Bloom, 1963). It has a blue-gray color when fresh, and an olive gray color when weathered. Fractures in the weathered Presumpscot Formation can have iron-manganese (?) stain along them. The weathered mud was previously considered a different unit from the blue-gray mud, deposited by different glacial advances (Trefethen and others, 1947). Leavitt and Perkins (1935), Goldthwait (1949, 1951), Caldwell (1959), and Bloom (1960, 1963), however, attributed the color difference to postglacial oxidation. Kelley (1989) and Mayer (1990) have discussed the mineralogy of the Presumpscot Formation from both offshore and onshore samples.

The Presumpscot Formation was deposited by glaciofluvial discharges directly into the glacial sea, the winnowed fine-grained fraction settling out as glaciomarine mud (processes discussed in Retelle and Bither, 1989). A sandy facies of the Presumpscot Formation is found overlying the fine-grained facies (Thompson, 1982, 1987). The contact between the facies may be sharp, although it is more commonly gradational. The informal term *sandy Presumpscot Formation* when used as a mappable unit (Weddle, 1987; Smith, 1988; Hildreth, 1990; Hunter, 1990; Retelle, 1991) is associated with marine regressive deposits, found stratigraphically above the massive mud of the Presumpscot Formation (*sensu stricto*). Alternatively, where the unit is thick enough to be separately mappable, the field trip leaders believe the term *nearshore deposit* is better suited for these shallow water or wave-reworked deposits associated with marine transgression and regression. To emphasize the reason for making this distinction, in some instances, massive sand of fluvial origin and unconformably overlying the Presumpscot Formation has been mapped incorrectly as sandy Presumpscot Formation.

Two symposia on the Presumpscot Formation provide an excellent overview of its geologic and engineering characteristics (cf. Andrews and others (1987), and Northeastern Section Geological Society of America Abstracts with Programs, 1988, SEPM symposium: "Glacial marine sedimentation: the Presumpscot Formation of northeastern North America and analogues").

Shallow Marine Deposits. Shallow or nearshore marine sediments were deposited after the ice margin and sediment source retreated from a depositional center, either as sea level overstepped the landforms, when isostatic sea level rise was greater than isostatic rebound, or as sea level fell around the landforms due to isostatic rebound exceeding eustatic sea level rise. The shallow marine facies (Smith, 1985; Retelle and Bither, 1989) contains a range of litho- and biofacies ranging from well-sorted tidal to subtidal sand to coarse bouldery lag deposits or lagoonal mud. The deposits include a variety of morphostratigraphic units such as beaches, spits, tombolos, subtidal sand bodies, and an extensive and nearly ubiquitous veneer of wave-reworked sediments. The deposits also display a wide range of textural maturity reflecting the source landform subjected to wave reworking and the available energy.

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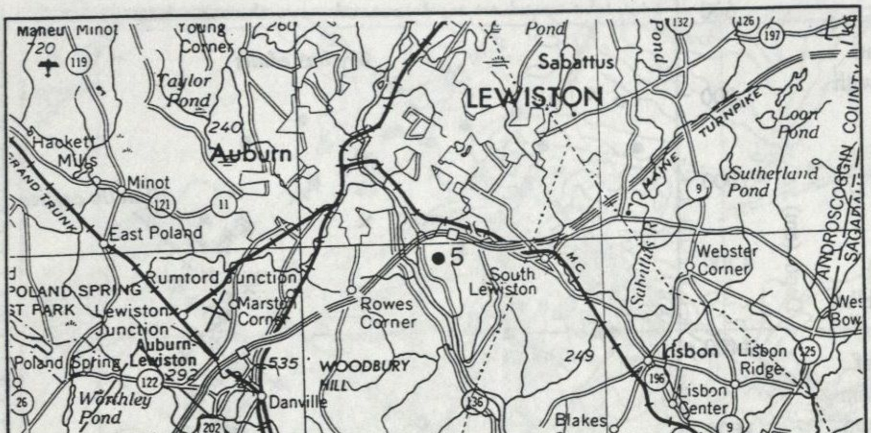

Figure 1. Location map and approximate sites of field trip stops.

ROAD LOG (Figure 1)

Mileage

0.0 **Assembly point; Saturday, October 7, 1995, 8:00AM.** Meet at parking lot of Topsham Fair Mall (McDonald's available). Location is east of Exit 24 off I-95; from Brunswick follow Rt. 24 (Maine Street) and cross Androscoggin River to Topsham; continue on Rt 196 to I-95 (Mall is on left just before I-95). *Park excess vehicles in commuter area in mall lot to minimize caravan; please carpool with vans as much as possible.* Leave mall, turn left at traffic lights, cross over I-95 and access I-95 southbound ramp (left). Proceed to exit 20.

9.5 Exit 20; left at stop sign at end of ram. Proceed to next stop sign 0.2 mile, turn left; sharp right 0.3 mile.

13.8 Libby Road (sign for Scott Dugas Knight Pit and Blueberry Pond Camping); turn right; proceed 0.1 mile; gravel pit on left, park on right side of road and walk into pit.

STOP 1. KNIGHT PIT. Park on Libby Road outside gate; 40 minutes; North Pownal quadrangle; elev. ca. 200 feet (60 m).

This stop is in a submarine ice-marginal moraine-fan complex, in which evidence for ice-shove and ice-contact collapse is well exposed. The maximum marine limit in the area as determined from deltaic topset/foreset contacts is about 300 feet (91 m) asl (Thompson and others, 1989). Articulated *Portlandia arctica* from a similar pit less than 5 miles to the east in the adjacent Lisbon Falls South quadrangle have dated at 14,045 ± 95 yr BP (AA10164; Weddle and others, 1993). The moraine ridge in the Knight pit is most likely of similar age. An AMS-radiocarbon age-date of 13,300 ± 50 yr B.P ($\delta^{13}C = -0.96$ ‰, OS-4419) on *Mytilus edulis* from nearshore deposits in the Knight Pit implies regression in the region was underway by that time. The mussel shells are found in Presumpscot Formation mud immediately underlying regressive sandy deposits; elsewhere in the pit, coarse nearshore deposits with boulder lag can be seen. The mussels appear to have been transported because some of the shells are found "spooned" or nested in one another, often in fragments, although articulated whole shells are common as well. The fossil-bearing mud has a coarse component, suggestive of redeposition as a debris flow from upslope. Foraminifers in the mud infer a polar to subpolar water temperature, and are typical of forams found in transgressive deposits elsewhere in Maine (C. Dorion, pers. comm., 1995; Cotter, 1985). Mussels are usually indicative of the intertidal and shallow subtidal zone, prefering rocky or pebbly bottoms, and where there is no large salinity variation.

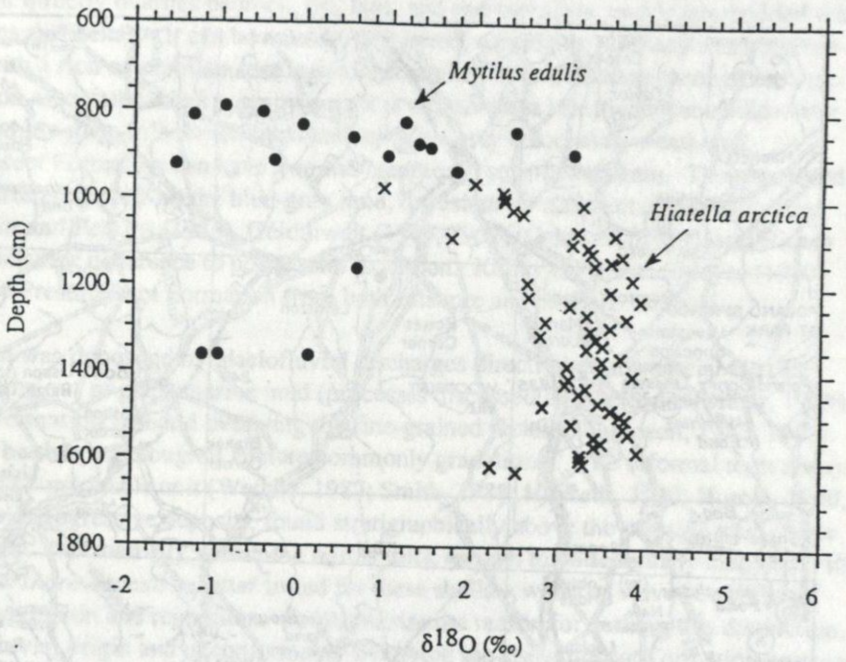

Figure 2. δ18O/depth, Lily Lake core, eastern Maine (depth cm below water surface), from Kreutz (1994).

Did the mussels live in the sea penecontemporaneously with the microfauna? As the assemblage and stratigraphy imply, the mussels were washed into place, most likely by storms. If so, the mussels are from a higher elevation, closer to the maximum marine limit in the area. The $\delta^{18}O$ values for the mussels were determined at the Stable Isotope Laboratory at the University of Maine. Serial sampling of growth bands on two shells gave a range of $\delta^{18}O$ values from -1.722 ‰ to 1.184 ‰ (avg. -0.648 ‰ and -0.037 ‰, n = 17).

Kreutz (1994a,b) measured isotopic variations on shells from several marine mollusc species from eastern Maine. At several sites, $\delta^{18}O$ values for *Mytilus edulis* averaged 0.9 ‰ (Pond Ridge); 1.27 ‰ (Sprague Neck); 1.45 ‰ and -0.67 ‰, (Lewis Cove, regressive and transgressive assemblages, respectively); and -0.26 ‰ (regressive assemblage, Lily Lake). Comparing the $\delta^{18}O$ values from the regressive assemblage to the transgressive assemblage of all species measured, Kreutz (1994) was able to show stratigraphically that the positive values of ice-proximal fauna remained unchanged upsection, suggesting little meltwater influence on shell $\delta^{18}O$ values until the regressional assemblage, where values decrease suggesting greater variability in both temperature and salinity during regression (Figure 2).

The importance of the $\delta^{18}O$ values of the mussels from the Knight Pit has yet to be fully evaluated. The elevation location and age of the mussels infer that they lived soon after the ice left the area, and at an early time during the regression (close to maximum marine limit). The average $\delta^{18}O$ values could reflect changing salinities due to glacial meltwater, meteoric water, sea ice melting in summer, or riverine discharge. Alternatively, the slightly negative values could be due to changing surface water temperatures during the growth season.

14.1 Return to vehicles, turn around in gate entrance and return to main road; turn right.
14.6 North Pownal center (stop sign and blinking red light); turn right onto Route 9.
17.9 Turn right at Big Skye Acres Campground site; turn into gate on left 0.2 mile, proceed to north end of pit.

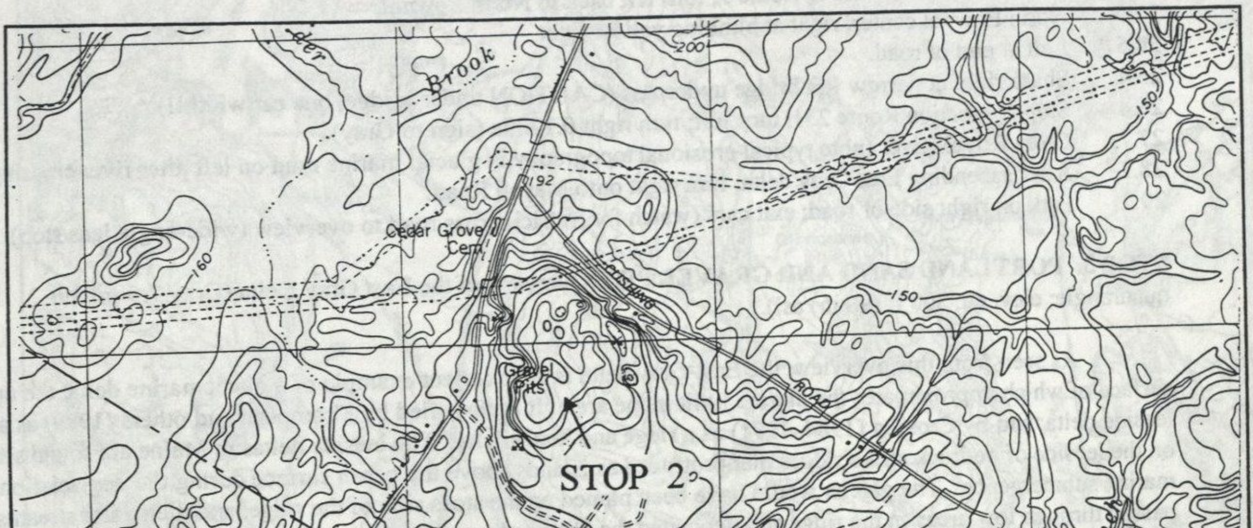

Figure 3. Stop 2 location map.

STOP 2. BLACKSTONE PIT. 40 minutes; North Pownal quadrangle; elev. ca. 300 feet (90 m).

The Blackstone gravel pit is located within a large glacial marine fan; it is connected to another fan to the south by a Pleistocene spit. Although the elevation of the top of the feature is just over 300 feet (90 m) (Figure 3), very close to the marine limit in the area, no deltaic topset / foreset contact is visible in the exposure. The deposit was formed during the retreat of the late Wisconsinan ice sheet at the ice margin. Foreset beds dominate the exposure, best seen in the central portion of the pit where they are uniformly dipping to the southwest. Presumpsco

Formation mud interfingers with sand beds along the southern, oceanward portion of the pit. On cleared faces, small-scale syndepositional faults and dewatering structures are present. Proximal and intermediate to distal facies of the glacial marine fan are represented by the coarse and fine sand and the interbedded fine sand and mud, respectively. In the northern end of the pit, a reworked sandy deposit overlies the sandy foreset beds, and represents wave reworking of the fan at the marine limit. A clast-rich unit of massive character unconformably overlies the fan and reworked beds, and is itself a reworked deposit formed during the marine regression. This unit is traceable down the topographic slope of the fan and consists of small to medium sized boulders within a matrix of fine sand to gravel. Although no apparent bedding is visible within the coarse-grained material, pockets of fine-grained material do contain some cross-bedding.

18.6	Return to vehicles and to Route 9; turn left back to North Pownal.
21.9	North Pownal center, right at blinking yellow light.
24.5	Left at end of road.
25.4	Sharp right at narrow RR bridge underpass (CAUTION under bridge, one car width!!).
26.4	Stop at junction Route 231, turn left; turn right 0.1 mile (sign to Gray).
27.7	Cross Royal River (note typical erosional topography in glacial marine mud on left after river crossing).
28.7	After ascending East Gray delta, turn right onto Mayall Road.
29.2	Park on right side of road; exit cars (watch for traffic); cross road to overview (wide-angle lens stop).

STOP 3. PORTLAND SAND AND GRAVEL PIT. Overview of the East Gray Delta (Gray 7.5-minute quadrangle; elev. ca. 300 ft (90 m) asl).

As seen from this overview, the East Gray delta is an excellent example of a glaciomarine delta, the upper surface of which approximates the marine limit in the area. It is classified by Thompson and others (1989) as a leeside delta, and by Crossen (1984, 1991) as a ridge and kettle delta. The leeside deltas in Maine are found situated on the leeside of bedrock strike ridges that protruded as islands above the ocean surface during the deglaciation and marine submergence. The glacier would have been pinned on the stoss side of the ridge, and meltwater streams passed through low areas in the ridge and deposited the delta sediment on the lee side of the ridge. At the East Gray delta, paths of meltwater streams cross the ridge that trends northeast from Gray Village (Thompson and others, 1989). A large kettle can be seen on the topographic map, occupying the central part of the delta (Figure 4). In context with leeside delta formation, the kettle is a result of a detached ice block, isolated from the main ice mass by the ridge, and is buried by the deltaic sediments and later melts and collapses the sediments (Crossen, 1984). A similar origin for deltas in the Belgrade Lakes region is described by Caldwell and others (1985).

The kettle and delta in the East Gray delta was reported by Stone (1899, p. 230), and by Leavitt and Perkins (1935) to be one of the largest kettle holes in Maine (a photograph of which is found in their report on page 98). Moreover, this site was visited by Perkins during the 1934 NEIGC, the second to be held in Maine, as reported in *Science* magazine (Reports, v. 80, p. 453, 1934). Later, the kettle was investigated by Goldthwait (1951) as part of the marine clay survey by the Maine Geological Survey. He reported that no clay was found in the kettle, and also that a wave cut scarp on the front of the delta demonstrated that the delta had not been submerged during the transgression. Crossen (1984) discusses the scarp on the East Gray delta along with scarps on other deltas in southwestern Maine in context with glacial isostatic rebound of the crust during deglaciation. She interprets the scarps to indicate that rebound during and following deglaciation uplifted early-formed deltas above eustatically rising sea level to produce the scarps, even as later deltas were forming at the upper marine limit. Southwest of the fieldtrip area, Koteff and others (1993) report wave-reworked features on the top of several deltas as evidence for submergence by rapid eustatic sea-level rise and a delayed postglacial rebound (cf., Thompson and others, 1989; Weddle and others, 1993).

The panorama of the pit operation provides a distant glimpse into the sedimentology of the delta. The northern portion of the pit is ice-proximal and contains huge boulders (some refrigerator-sized clasts are present), along with a generally coarser component than the southern part of the pit, where the topset/foreset contact of the delta is exposed. The foreset lobes of the delta dip in a wide trend, from northeasterly to southerly in different parts

Figure 4. Location map for Stops 3 and 4, and East Gray and Gray deltas.

of the pit. The remnants of the kettle are visible in the central part of the pit, however, most of that feature has been mined away. In the photograph from Leavitt and Perkins (1935), the bottom of the kettle is shown prior to excavation. Large boulders are visible in the photo, and a small ridge described in the photo caption as an esker (also visible on the quadrangle) can be seen separating the two large depressions of the kettle. This ridge trends southerly to the part of the delta that is south of the large kettle, and is mantled by the delta or terminates at that position. Older deltaic deposits are present southwest of the East Gray delta toward Gray Meadows, and are apparently associated with the glacier drainage system the ridge is part of. The configuration of the esker (?) ridge and kettle suggests that the East Gray delta is a composite delta, where its southern portion has an ice-contact origin, and the northern part of the delta (north of the kettle) is the leeside component as described above.

In an exposure on the northeast side of the delta in Collyer Brook valley (Figure 4), distal deltaic bottomset beds, once continuous across the valley, are exposed. They have been eroded by late- and postglacial drainage producing the morphology in Collyer Brook valley that is typically found in areas underlain by fine-grained glacial marine deposits in Maine. Bloom (1960) described a 53-foot thick section along Collyer Brook, about 0.5 mile east of the delta front, at which most of the section is comprised of Presumpscot Formation, gradationally overlain by thin sand and mud beds and eventually capped by fine to coarse sand. He reported marine fossils are found near its base, including *Nucula expansa* and *Portlandia arctica*. However, this section is overgrown now, and no fossils could be found here during a visit to the site. The East Gray delta probably was deposited about14,000 radiocarbon yr B.P., similar to the Florida Lake end moraine complex about 10 miles east in Freeport, from which a *Portlandia arctica* dated at 14,045 ± 95 yr BP (AA10164; Weddle and others, 1993; Lisbon Falls South quadrangle, in press).

The East Gray delta is infamous because it is the site of the McKin Superfund site, one of the earliest identified locations of this designation. Trip C-7 (Hydrogeology and Environmental Geology of the Gray Deltas) in this guidebook will focus on this aspect of the delta.

31.4 Return to vehicles and continue straight (view ice-proximal end of delta as we drive by); proceed to stop sign. Cross Lewiston Road (Route 202/100/4) with caution.

31.8 Begin ascent of Gray delta front; top of Gray delta 0.2 mile (potato field on left).

The Gray delta (Figure 4) has been described by Bloom (1960) as a "proglacial marine delta, in which a complete transition from ice-contact stratified drift through outwash and deltaic sand into Presumpscot Formation can be traced. ......The late-glacial and post-glacial history of the vicinity of Gray is not complex. In brief, a group of rock hills were fringed with kame terraces when glacier ice was still present on the valley floors. Slightly later the valleys were submerged by the sea nearly to the tops of the kame terraces and became branches of an estuary. At that time, still-melting ice in and northwest of the Sabbathday Lake basin discharged meltwater into the estuary, and a flat topped delta of sand was built out over the marine mud. Subsequent emergence has resulted in minute dissection of the impermeable, clay-rich marine sediment, while the permeable outwash and ice-contact stratified drift have retained their constructional topography."

Thirty-five years later, the deglacial history is interpreted somewhat differently (Crossen, 1984; Thompson and others, 1989; Weddle, mapping in progress). The delta marks the maximum marine limit in the area, and is comprised of coalesced ice-contact or lee-side deltas (heads at Sabbathday Pond and Crystal Lake, respectively), with discharge to a relative falling sea level (evidenced by meltwater channels on its surface). The deltas and the Presumpscot Formation are deposited penecontemporaneously, and are interbedded in the distal part of the deltas. A radiocarbon age-date on bulk material is reported from cores from Sinkhole Pond on the delta top, and from Poland Spring Pond just to the north (12,710 ± 125 (SI-4657), and 12,860 ± 325 (SI-4656), respectively; Davis and Jacobson, 1985). These dates are close in age to recently reported dates from marine shells in the Androscoggin River valley to the east (Weddle and others, 1993; see Stop 5 this trip). However, in both cases, the terrestrial vegetation and the marine shell dates are minimum limiting dates, and do not necessarily date the time the ice margin was in the area. Other work in the area includes compilation of test-boring logs of the Maine Turnpike Authority by Prescott (1980), and studies by private geological consultants on the ice-contact deltas and fans south of the Gray delta (discussed in Trip C-7, this guidebook).

33.3 Cross over Maine Turnpike; proceed on road, parallel to meltwater channel on left. We will descend into and out of the channel and back up onto the delta surface farther up the road.
34.7 Stop sign, turn right and continue, staying on Route 26.
35.4 Sabbathday Pond on right; ice-contact head of Gray delta.
37.0 Shaker Village; turn right into lot and park where available space allows.

LUNCH STOP. THE SHAKER MUSEUM, NEW GLOUCESTER. (Lunch not provided; bring your own!! One hour) Situated on Sabbathday Pond, the Shaker museum is a living museum of America's oldest religious community. Field trip participants may visit the Shaker Store (gift shop), the Shaker Museum Reception Center (bookshop and exhibits), and the 1816 Spinhouse (exhibit of late era Shaker furniture).

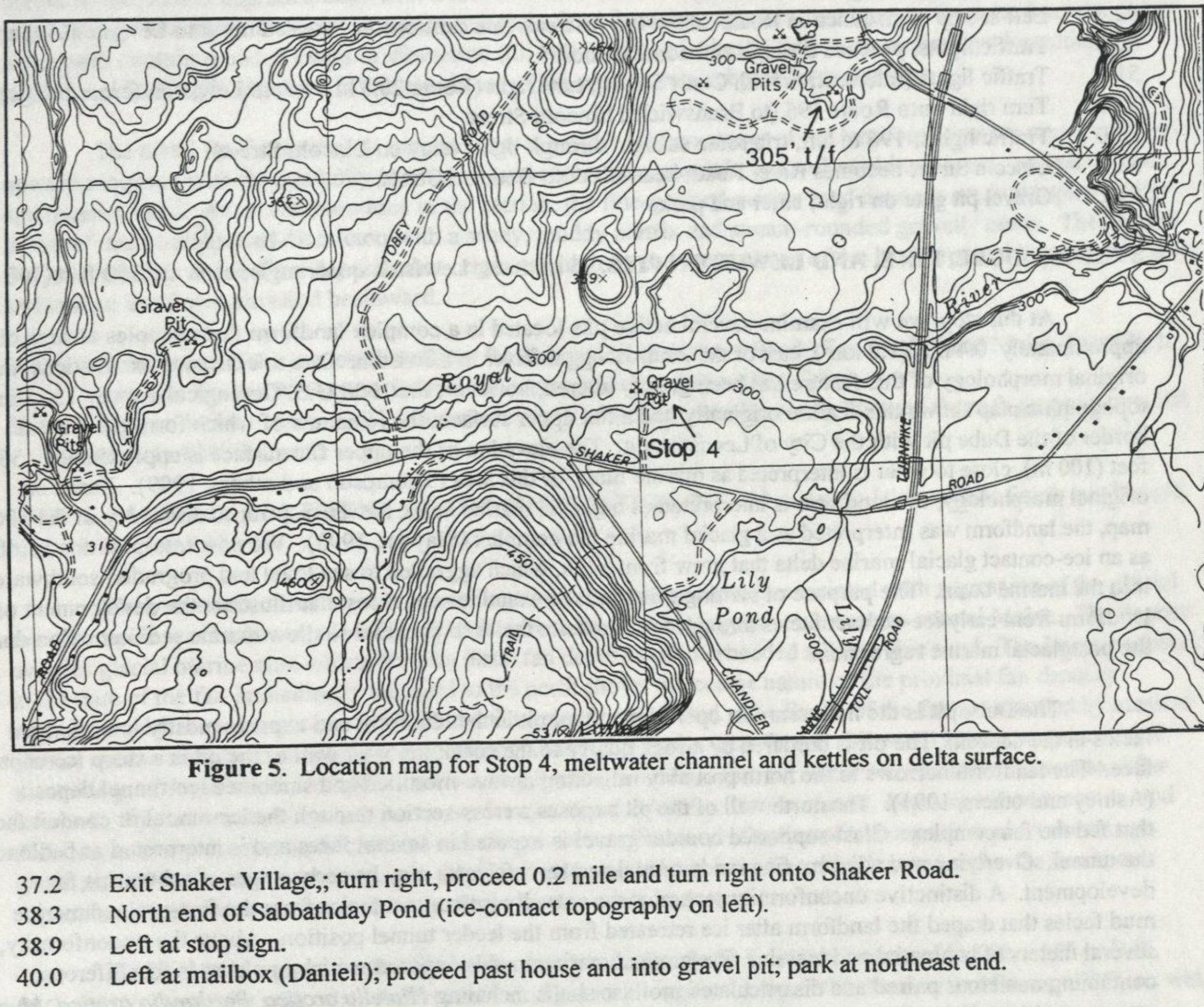

Figure 5. Location map for Stop 4, meltwater channel and kettles on delta surface.

37.2 Exit Shaker Village,; turn right, proceed 0.2 mile and turn right onto Shaker Road.
38.3 North end of Sabbathday Pond (ice-contact topography on left).
38.9 Left at stop sign.
40.0 Left at mailbox (Danieli); proceed past house and into gravel pit; park at northeast end.

STOP 4. TOWN OF NEW GLOUCESTER PIT. 40 minutes, Gray quadrangle; elev. ca. 285 feet (86 m).

This gravel pit is located in a meltwater channel incised into a delta surface (Figure 5). Numerous kettles are present on this surface, and collapsed beds associated with a melting ice block in the meltwater channel are visible in the east wall of the pit. The shallow pit exposes fluvial deposits over deltaic foresets(?). However, the elevation at this site is too low to be a topset/foreset contact associated with the maximum marine limit. Thompson and others (1989) group this delta with the Gray delta (their name is Crystal Lake-Sabbathday Pond delta). A topset/foreset contact at 305 feet (92 m) asl was measured in a pit north of this stop (pers. comm., W. B. Thompson, 1994). This delta was deposited concurrently with or soon after the Gray delta. The stagnant-ice margin was to the west at the north end of Sabbathday Pond, and remained there some time after the delta formed, with meltwater streams incising the delta surface as relative sea-level fell. Extensive ice-contact topography north of Sabbathday Pond in the Poland Spring area reflect this zone of stagnant ice at the active glacier margin. Test-boring data along the Maine Turnpike north of the incised delta record at lower elevations thick sand over marine deposits. The sand most likely represents redistributed sand from the incised delta, deposited at lower elevations during relative sea-level fall.

40.7	Return to vehicles, return to road and turn left.
42.1	Yield sign (bear right with caution).
42.4	Left at stop sign, Junction Route 202/4/100 (use extreme caution!!). Proceed north to Lewiston/Auburn.
50.6	Traffic lights, proceed through and continue north.
51.1	Traffic lights (intersection with Court Street), turn right (Route 202/11/100) to Lewiston (cross bridge).
51.6	Turn right onto Route 196 (to Brunswick); Lincoln Street.
52.0	Traffic lights; 196 to left, trip route straight through lights (stay on Lincoln Street).
53.7	Lincoln Street becomes River Road; bear left onto Goddard Street.
54.2	Gravel pit gate on right; enter and park.

STOP 5A. DUBE, PIKE, AND LEWISTON PITS. 75 minutes; Lewiston quadrangle; elev. ca. 330 feet (100 m).

At this stop, we will examine several active pits located in a complex landform that occupies an area of approximately 0.4 mi^2 (1.0 km^2), east of the Androscoggin River in Lewiston. Due to extensive excavations, the original morphology of this feature has been greatly altered, however, the 1908 U.S. Geological Survey 15-minute topographic map shows this feature originally had a flat upper surface, the remnants of which form the current border of the Dube pit with the City of Lewiston pit. The elevation of the upper flat surface is approximately 330 feet (100 m), close to what is interpreted as marine limit for this area (Thompson and others, 1989). Based on original morphology, the landform is interpreted as a glacial marine delta, however, prior to discovery of the 1908 map, the landform was interpreted as a glacial marine fan complex (Slayton, 1993). We now interpret this landform as an ice-contact glacial marine delta that grew from a fan, which aggraded to sea level and prograded southward into the marine basin. The purpose of visiting this site is to examine sequences that illustrate the development of the landform from early ice-contact facies through fan and delta facies to eventual shallow marine sedimentation during the postglacial marine regression.

The Dube pit is the northernmost operating gravel pit in the landform and exposes mostly ice-proximal facies in the deposit. The pit is bordered by a rock quarry to the north; the west wall of the pit is a steep ice-contact face. The landform narrows to the north probably reflecting a wave-modified and smoothed ice-tunnel deposit (Ashley and others, 1991). The north wall of the pit exposes a cross-section through the ice-tunnel or conduit facies that fed the fan complex. Clast-supported boulder gravel is exposed in several faces and is interpreted as bedload in the tunnel. Overlying and adjacent fine sandy gravel and sand foresets may be early stages of submarine fan development. A distinctive unconformity separates the gravelly ice-contact facies from the fine-grained marine mud facies that draped the landform after ice retreated from the feeder tunnel position. Above the unconformity, several meters of laminated and massive fine-grained marine mud is exposed, which in places is fossiliferous, containing numerous paired and disarticulated mollusc shells including *Hiatella arctica, Portlandia arctica, Mya truncata,* and *Mytilus edulis,* as well as *Balanus* plates. *Portlandia* shells recently yielded an age of 12,980 ± 85 yr BP (AA10165), and may be interpreted as dating the ice margin at this site. There is, however, evidence of reworking in the deposits as the fossils at this site are found in several contexts, including both *in situ* in the massive mud, but more commonly in a coarse-grained diamicton that is exposed between the laminated mud and the east wall of the pit. The diamicton is interpreted as a debris flow deposited down the dipping foreset beds of the delta. The marine mud is capped by sandy and muddy offlap sediments whose structures mirror the underlying topography of the landform.

54.8 Return to River Road; turn left at stop sign (watch for traffic with caution!).

55.8 Park at Dragon Products buildings (right side of street); exit vehicles with caution and cross road to pit entrance (watch for traffic).

STOP 5B. DUBE, PIKE, AND LEWISTON PITS (Continuation of Stop 5A).

There are several exposures in the central and southern part of the landform (Lewiston and Pike pits) that nicely illustrate the sedimentary regimes in the medial and distal portions of the complex. A north-south oriented wall in the central area of the pit exposes thin south-dipping sandy foreset beds, comprised predominantly of normally graded and ungraded beds with a few 4 - 8 inch (10 - 20 cm) thick beds of ripple-drift cross laminated sand. The graded sandy beds are interpreted as grain flow deposits originating upslope on the delta foresets. Some sandy beds contain rounded clasts of diamicton (till balls) up to 2 inches (5 cm) in diameter. Syndepositional extensional faulting is also seen in translocations of the sandy strata subparallel to bedding.

The distal fan facies is well exposed near the southwestern corner of the landform near the top of a fining-upward sequence with rhythmically laminated sand at the base, overlain by sand-mud interlayers, laminated mud, and massive sand. At several locations in the section, the fine-grained marine mud deposits are interrupted by thin layers of matrix-supported diamicton with a sandy, muddy matrix and stream-rounded gravelly clasts. These units are interpreted as debris flow deposits that originated high on the foreset beds of the landform and incorporated sand and mud as the flows travelled basinward.

56.0 Return to vehicles, proceed on River Road; turn left at Armory Road (sign for 495 and Maine Turnpike).
57.4 Junction Route 196 (east to Brunswick, through Lisbon and Lisbon Falls).
69.1 Sign on left for Maine DOT Maintenance lot); turn right onto dirt road. Proceed through gate and turn left to end of pit and park.

STOP 6. MAINE DEPARTMENT OF TRANSPORTATION PIT. 30 minutes; Lisbon Falls South quadrangle; elev. ca. 200 feet (60 m).

The purpose of this stop is to see a Pleistocene nearshore deposit associated with regression of the glacial sea after uplift was underway, and discuss radiocarbon-age chronology in southwestern coastal Maine. The deposit overlies glacial marine mud which rests on distal fan deposits of interbedded silt and fine sand. The larger exposure in the core of the fan immediately adjacent to the north shows the coarse nature of the proximal fan deposits associated with the ice tunnel from which the fan deposits originated. The flanks of the fan are mantled by massive glacial marine mud from which the pit operator claims fossil shells were found. Bloom (1960) reported a varied assemblage of fossil shells from this pit including *Hiatella arctica, Macoma calcarea, Musculus substriata, Mya arenaria, Mytilus edulis, Nuculana jacksoni, Serripes groenlandicus, Natica clausa, Neptunea decemcostata,* and *Balanus,* and represent mixed intertidal and deeper-water affinities, reflective of the depositional environments associated with the sediments in the pit. Molds of pelecypods have been found in the mud under the nearshore deposit but fossils have not been recently observed elsewhere in the pit.

Smith (1985) reports a radiocarbon age-date on marine shells from this location (11,650 ± 175 yr B.P., DIC-1501). In context with other earlier reported dates in the region (Attig, 1975; Stuiver and Borns, 1975), as well as the most recently reported dates noted in this guide, these dates bracket the marine regression to between ca. 13,300 to 11,500 yr B.P. This duration span for the regression at this site is problematical for at least two reasons. Although Smith (1985) reported the date from this stop, he did not report the elevation of the sample site. In fact, the surface elevation here is close to the elevation the 13,300 yr B.P. age-date came from. Moreover, the dates reported by Retelle and Bither (1989) and Weddle (1994) on shells associated with regressive deposits at about 12,800 yr B.P. are at much lower elevations than this stop. Dates from wood and from marine shells from the Presumpscot Formation near sea level near Portland suggest the regression had reached that elevation sometime between about 11,900 and 11,500 yr B.P. (Anderson and others, 1993; pers. comm., W. B. Thompson, 195). The sample Smith (1985) reported from the DOT pit (this stop) was not a large amount of material and was not leached during the lab preparation (pers. comm., G. W. Smith, 1995), and hence may have contamination error.

END OF TRIP; return to Route 196, turn right and proceed to Topsham Fair Mall (on right after crossing I-95). Pick-up vehicles and leave mall; left to I-95, right to Brunswick and Bowdoin College.

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