Title: Late Wisconsinan Glacial and Marine Geology and Early Postglacial Geomorphic Evolution of the Lower Androscoggin Valley and Casco Bay Lowland

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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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LATE WISCONSINAN GLACIAL AND MARINE GEOLOGY AND EARLY POSTGLACIAL GEOMORPHIC EVOLUTION OF THE LOWER ANDROSCOGGIN VALLEY AND CASCO BAY LOWLAND

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INTRODUCTION

Our current understanding of the setting, mechanisms, and timing of ice marginal retreat in the modern glacialmarine realm (e.g. Trusel et al., 2010) provides a basis for interpreting morphostratigraphic evidence of dynamic Pleistocene age marine-based glaciers and their connection with sea level change (Powell and Cooper, 2002; Corner, 2006). Conversely, a more thorough understanding of the history of dynamic ice sheets gives valuable insights to the future behavior of marine ending glaciers that are predicted to appreciably increase sea level rise in the next century (Mouginot et al., 2015).

The surficial geology of the coastal lowland and western foothills of southwestern Maine records the systematic recession of a marine-based sector of the late Wisconsinan Laurentide Ice Sheet. On this field trip we will examine landforms and exposures in the lower Androscoggin Valley and adjacent Casco Bay Lowlands (Figure 1) where assemblages of ice-marginal features lie in contact with glacialmarine deposits. These deposits show that an active, grounded ice margin retreated in contact with a synglacial sea in a glacioisostatically depressed foreland in front of the ice sheet. Raised marine deposits in the lowlands are associated with the marine transgression and subsequent regression, recording the interplay between glacioisostatic and eustatic sea level change. We will also examine deposits that represent the shift in the mode of ice recession from active retreat to stagnation zone retreat which occur at the transition between the coastal lowland and the western foothills.

The Field Area

The lower Androscoggin Valley and Casco Bay Lowland study area extends from the present Maine coast to the foothills of the White Mountains in the northwest and is bordered by drift covered bedrock terrain to the northeast and southwest. Denny (1982) and Hanson and Caldwell (1989) previously defined the Coastal Lowland and Central Upland as the principle physiographic units in the region. Total relief in the lowland is generally less than 150 meters while in the foothills of the Central Uplands the relief can exceed 300 meters.

The bedrock in the lowlands is comprised of northeast-striking Silurian-Ordovician age metasedimentary rocks locally intruded by Devonian/Carboniferous-age plutonic rocks (Berry and Hussey, 1998; Hussey and Berry, 2003). The coastal metamorphic terrain is bordered to the west by the Carboniferous age Sebago Batholith. Bedrock is best exposed in the uplands and along the immediate coast while in the lowland areas, bedrock is generally less well exposed due to blanketing by surficial sediments. The region is drained (from east to west) by the Cathance, Sabattus, Androscoggin, Royal and Piscataqua Rivers.

LATE WISCONSINAN GLACIAL GEOLOGY AND SEA LEVEL HISTORY

At its most recent maximum advance during the late Wisconsinan stage, (marine isotope Stage 2) the southeastern margin of the Laurentide Ice Sheet extended across the Gulf of Maine onto the continental shelf on George’s Bank. (Tucholke and Hollister, 1973; Stone and Borns, 1986; Schnitker et al., 2001). Retreat from George’s Bank likely began as early as 22,000 ¹⁴C years B.P. but no later than 17,500 B.P. (Schnitker et al, 2001). The ice configuration and style of deglaciation from the LGM position has been a source of debate. Hughes et al., (1985) and Hughes (1987) proposed that inland ice in the foothills and coastal lowlands fed an ice stream which, in turn, fed a floating shelf over several of the deeper basins of the Gulf. They further suggest that by ca. 14,000 B.P., the Gulf of Maine became a calving embayment. A similar model, based on the interpretation of
sediment core and seismic data led Belknap and Shipp (1991) and Schnitker et al., (2001) to propose that ice was initially grounded in the Gulf during the LGM however in an early phase of the retreat “temperate” ice shelves were formed over the deeper basins in the Gulf by marine waters entering under the ice margin. Oldale et al., (1990) proposed an alternative model for a grounded marine ice sheet that retreated across the Gulf, similar to modern tidewater glaciers in Alaska and Svalbard.

Figure 1. Map of the lower Androscoggin Valley-Casco Bay Lowland. Field trips stops are numbered. Letter A is the assembly point; 4/L is the lunch stop. Inset map shows location of field area. Shaded area on the inset map illustrates the area below the late Wisconsinan marine limit (base map from Google Maps).

It is generally agreed that the retreating ice margin was grounded when it reached the present coast south of Portland. Progressive ice marginal retreat through the Androscoggin Valley and Casco Bay Lowlands is marked by ice-marginal deposits interbedded with marine sediments, deposited as the synglacial sea transgressed across the isostatically depressed foreland south of the ice margin. The retreat through the lowland was punctuated by minor, and perhaps some longer, still-stands represented by moraines, grounding line fans, and glacialmarine deltas described below.

**Synglacial and Postglacial Sea**

The inland and vertical extent of the synglacial sea is defined by raised shoreline features and glacialmarine deltas that increase in elevation from the coast to the interior in response to the isostatic load of the Laurentide Ice Sheet (Thompson et al., 1989). Marine limit increases from around 75 m along the outer coast near Phippsburg to over 100 m in the Androscoggin River valley north of Lewiston (Thompson et al., 1989). The marine limit was established synchronously with ice retreat by the transgressing sea; thus the oldest marine sediments across the lowland range in age from ca. 14,000 to 13,000 yrs B.P. The marine transgression was quickly followed by regression of the synglacial sea as relative sea level dropped due to rapid postglacial isostatic rebound (Bloom, 1960, 1963; Stuiver and Borns, 1975; Barnhardt et al., 1995). An AMS-radiocarbon age on Mytilus edulis from a nearshore deposit at 60 m asl in Pownal infers that regression in the region was underway by 13,300 ± 50 yrs B.P.
In Topsham shell ages from intertidal deposits indicate that the inland sea had regressed to near 30 m asl by approximately 12,800 yrs B.P. (12,820 ± 120 yrs B.P. (SI-7017), Retelle and Bither, 1989; 12,850 ± 45 yrs B.P. (OS-2348), Weddle, 1994). The Brunswick Sand Plain, an extensive regressive delta graded to a sea level at approximately 18 m asl at the mouth of the Kennebec-Androscoggin Valley between 12,500 and 12,000 yrs B.P. (Retelle and Weddle 2001). Sea level continued to drop to the postglacial lowstand on the inner continental shelf at a depth of approximately -55 m by 10,500 yrs B.P. (Barnhardt et al., 1997).

Geochronology

The timing of ice retreat and sea level history in this region is constrained by radiocarbon dated materials, predominantly marine shells that were found in close association with ice marginal positions or shorelines. In recent field trip guides (e.g. Weddle and Retelle, 1995) and in Retelle and Weddle (2001) we presented uncorrected ages in radiocarbon years before present (14C yrs B.P.). In this guide we again report the uncorrected ages, however in the text below and in the field, we will discuss the significance of calibrating these ages as they apply to significant landforms that record ice marginal positions and sea level events related to independently dated paleoclimate records such as ice cores. It is, however, important to recognize that a marine reservoir correction needs to be applied for marine shells and that such a correction likely varied spatially and with time through the late Pleistocene. A normal starting point for marine reservoir correction is -400 years however there is significant regional variability across the North Atlantic region (Mangerud and Gulliksen, 1975; Bondevik et al., 1999; Coulthard et al., 2010). In eastern Maine, Dorion et al., (2001) compared basal organic lake sediments and marine sediments and reported a difference of -600 to -800 years. In the Portland area, Thompson et al., (2011) obtained paired ages on spruce logs and marine shells (*Mytilus edulis* and *Balanus sp*.) which indicated an offset of approximately 1,000 years. Application of their suggested 1,000 year correction allowed more direct correlation with the ice retreat history in central and western New England reconstructed using the New England varve chronology (Ridge, 2004, Ridge et al., 2012).

**GLACIAL AND GLACIALMARINE LANDFORMS AND SEDIMENTS**

**End Moraines**

Many moraine ridges have been identified and mapped in the coastal lowlands in Maine (Smith, 1982, 1985; Thompson, 2015). The characteristic morphology of the deposits are as linear features that are inferred to be aligned along the former ice margin. Moraines vary in size and composition. Smaller moraines may have as little as a meter or two in relief and a few meters wide while larger moraines may be 20 or 30 meters high and 100 meters wide (Thompson, 2015). Smaller moraines and clusters of these landforms in the lowland have been referred to as *DeGeer Moraines*. Examples of these moraines have been described in areas where ice sheets are retreating in contact with marine waters (e.g. Lindén, and Möller, 2005; Solheim and Elverhøi, 1997; Zilliacus, 1989, 2013). Larger moraines, such as the Pineo Ridge Moraine Complex and the Addison Moraine have been called *Stratified End Moraines* (Ashley et al., 1991) and these are commonly comprised of complexly bedded units of sand and gravel with interbedded lenses of diamicton. The internal structure of the stratified end moraines is commonly glaciectonized showing folds and faults from ice shove and overriding (Retelle and Bither, 1989).

**Ice contact deposits**

Sediment deposited by meltwater occurs in several genetically connected 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.
Glacial marine fans
Sediments delivered to the ice margin by glacial meltwater from sub- and englacial conduits and ice marginal streams are deposited at the grounding line in mound or cone-shaped accumulations that commonly drape or flank the moraine ridges. These deposits have been called glaciomarine fans (Sharpe, 1987), grounding line fans (Powell, 1990) or submarine outwash (Rust, 1977). In ice-proximal areas of the fans, situated at or near the mouth of the meltwater conduit, fans are typically comprised of coarse-grained gravel. Sediments fine distally away from the conduit. Fan sediments represent the transition between fluvially dominated processes associated with the ice tunnel environment, and processes of the proglacial marine basin. Consequently, fan sediments show complex stratigraphic relationships both parallel to the ice margin and distally from the former tunnel mouth. Sediments in the proximal zone include stratified coarse-grained sand and gravel supplied by the meltwater currents and diamicton that may have originated from slope failure and downslope movement from the adjacent moraine, or implacement by ice shove (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 laminated sediments, termed cylopsams and cyclopels (Mackiewicz and others, 1984) were likely deposited by suspension from a highly turbid but buoyant suspended sediment plume (Cowan and Powell, 1990). 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 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).

Glacialmarine Deltas
The origin and stratigraphy of glacialmarine deltas in Maine has been described 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 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 elevation of the contact 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 in deltas across the state to define the plane of the upper marine limit. Topset/foreset contacts increase in elevation from southeast to northwest, representing the increased isostatic loading and associated increased postglacial rebound. 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).

Fine-grained glacialmarine sediment
Fine-grained glacialmarine sediments that are found extensively across the Maine coastal lowlands were named the Presumpscot Formation by Bloom (1960). 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. The fine-grained sediment can be massive or laminated, containing outsized clasts and icerafted debris, and is fossiliferous, with a rich assemblage ranging from forams, diatoms, mollusc shells and marine vertebrates. Based on associated fossil assemblages, the Presumpscot Formation is characterized as a cold-water marine unit (Bloom, 1960, 1963). The sediment has a blue-gray color when fresh, and an olive gray color when weathered. The Presumpscot Formation was deposited by glaciofluvial discharge directly into the glacial sea, the winnowed fine-grained fraction settling out as glaciomarine mud (Retelle and Bither, 1989).
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 (Koteff et al., 1993), or as relative 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-rewilded sediments. The deposits also display a wide range of textural maturity reflecting the source landform subjected to wave reworking and the available energy.

Distal outwash deltas are formed where meltwater streams entered the sea some distance from the ice margin (Thompson et al., 1989; Tary et al., 2001). Isostatic uplift was in progress across much of the region as the ice withdrew inland from the marine limit. Broad sand plains are present in Maine and are commonly found at the inland extent of the marine limit and some of these distal deltas are associated with ice-marginal deposits (Thompson et al., 1989). The sand plains range in elevation from 136 m asl to 36 m asl (Weddle and Retelle, 1995). Some distal deltas are found well seaward of the inland marine limit. Of these, the Brunswick Sand Plain (BSP, Figure 9) is the farthest from the inland marine limit, and is found at the lowest elevation (Weddle, 1994; Weddle and Retelle, 1995). The surface of the BSP grades from about 36 m to 18 m asl (surface gradient 1.8 m/km). 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 this extensive landform as a coastal braid-plain delta (Nemec, 1990) formed as braided streams in the late glacial Androscoggin River valley entered the regressive sea. The braided-stream deposits unconformably overlie very low angle foreset beds. The foresets overlie a coarsening-upward sequence of massive and laminated mud grading up-ward to interbedded sand and glaciomarine mud, described previously as the sandy Presumpscot Formation. This muddy unit beneath the braid-plain delta represents a transition 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 sediments were deposited.

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Figure 2 from Retelle and Weddle (2001) shows the location and orientation of ice marginal deposits. In general, the individual deposits collectively show an east-northeasterly orientation inferring ice retreat in a north-northwesterly direction. The Cox Pinnacle Moraine is labeled as CPM.
ROAD LOG

Assembly Point: Dick’s Sporting Goods Parking Lot, Topsham Fair Mall Road, off Rte.196 in Topsham Maine. 43°55’59.18”N; 69°58’58.84”W. We will assemble in the lower section of the lot starting at 8:30 A.M. and leave at 9:00 A.M. sharp. Please carpool and consolidate in vans if possible. It would more convenient to bring your lunch with you. The lunch stop at the New Gloucester Fish Hatchery is not in town or close to convenience stores.

The deposits in the area have been mapped at the 7.5 minute quadrangle scale (1:24,000) as part of the Marine Geological Survey U.S. Geological Survey COGEOGRAPH surficial geology mapping program. The field trip will visit localities in the Brunswick, Freeport, Lisbon Falls South, North Pownal, Gray, Yarmouth, and Cumberland Center Quadrangles. The surficial geologic maps can be downloaded as pdf’s on the Maine Geological Survey website: http://www.maine.gov/dacf/mgs/pubs/online/surficial/surficial.htm

Mileage

0.0 Exit parking lot and follow Topsham Fair Mall Road to Rte. 196. Get in the left lane at the stoplight, turn left and follow Rte. 196 over Rte. 295. Continue ahead (Jack’s Pit 3.8 miles) on 196 to Lisbon Falls village.
6.5 Stoplight in Lisbon Falls with mill on the left in front of you, world famous Moxie Store across the street to the right). Turn left and follow the road across the river. Take a left at the T-intersection (Route 125). Stay on Rte. 125 until the powerline crossing at the top of the hill.
10.5 Parking on the right. Pull off as far as possible. Exercise caution in crossing the road!

STOP 1: COX PINNACLE MORaine,
Lisbon Falls 7.5 minute quadrangle, 43°56’48.6”N; 70°04’10.2”W

The Cox Pinnacle moraine (CPM in Fig 2) is larger and more continuous over a greater distance than most other end moraines in southwestern Maine, with the exception of the Waldoboro Moraine (Thompson and Borns, 1985). The gently lobate form can be traced for ~6 km across the landscape and delineates the ice-marginal position where the retreating ice margin was pinned on the bedrock up-land between the Androscoggin River and Royal River valleys. Sandy till is exposed in the drainage ditch alongside Rte. 125. On the upland, the moraine is composed of sandy diamicton and large boulders are present along its crest. This moraine is exposed in gravel pits to the east in the lowland of the Androscoggin River valley, where it is composed of till and deformed subaqueous outwash deposits (Weddle. 1997). The moraine is laterally correlative to other smaller moraine ridges and ice-contact features to the west and to a large fan and moraine complex to the east in Topsham (Retelle and Bither, 1989). Although the exact timing of emplacement of the moraine is not known at present, ice retreat from the grounding line at Topsham is older than 13,300 yrs B.P.

Return to the vehicles and again exercise caution crossing Rte. 125. Continue on Rte. 125 south to junction with Rte. 136 in Freeport

16.1 Turn left onto Rte.136. Continue ahead to the left turn where Rte. 136 continues over Rte. 295.
16.8 Turn right on Durham/Pownal Road before the turn to the Rte 295. Continue on Pownal Road until the right turn onto Libby Road
20.7 Turn right onto Libby Road. The entrance to the Knight Pit is on the left 0.1 miles ahead.
Figure 3. LIDAR image of end moraine topography in the Lisbon Falls South Quadrangle west of the Androscoggin River in Durham. The Cox Pinnacle Moraine is a major linear feature trending northeast to southwest from the river before it bends to a west-northwesterly trend delineating the margin of the Androscoggin lobe. Numerous minor moraines, here referred to as DeGeer moraines, occur both south and north of the Cox Pinnacle moraine.

STOP 2: KNIGHT PIT, SCOTT DUGAS EXCAVATION
North Pownal Quadrangle 43°53′27.5″N; 70°10′22.2″W; elevation 190 feet at top of landform)
This stop is in a submarine ice-marginal moraine-fan complex (Marvinney, 1999) in which evidence for ice-shove and ice-contact collapse has been 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 of 13,300 ± 50 yr B.P. 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. In the adjacent Blackstone Pit, steeply dipping fan foresets are overlain by a wedge of fine-grained Presumpscot Formation marine mud, presumably deposits after ice retreated from the grounding line. Cobbles and boulders in the marine mud may have been deposited by ice rafting or emplaced by gravity driven flows.
Figure 4. East central sector of the North Pownal 7.5 minute surficial geologic map Knight Pit (Stop 2) is located in the landform mapped as Pempc, Pownal Center End Moraine complex (Marvinney, 1999).

20.9 After leaving the pit turn around on Libby Road and take a right turn at the bottom of the hill. Follow Pownal Road to stop sign at the intersection of Rte 9. Continue ahead to Allen Road.

24.0 Take a left and follow Allen Road west to the intersection with Rte 231 at Pineland Farms.

25.9 Turn left onto Rte 231 at Pineland Farms and take a quick right (26.0 miles). You will drop steeply into the Royal River valley which is deeply incised into the glacialmarine mud. Several now-defunct brickyards which utilized the marine clay are located in this valley.

27.2 Cross the railroad tracks and Royal River. You will rise up onto incised seafloor deposits. Note the characteristic topography on the left at the crest of the hill.

28.1 Continue ahead to junction with Mayall Road (28.1 miles) and turn right at the stop sign. Follow Mayall road until you see the sign for Portland Sand and Gravel on the left.

28.6 Turn left into Portland Sand and Gravel Co.

STOP 3: EAST GRAY DELTA
Gray 7.5-minute quadrangle; elev. ca. 300 ft (90 m) asl). 43°53’37.6”N, 70°18’9.1”W

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 with a topset/foreset contact measured at 289 feet (88 m). 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 et al., 1989). A large kettle can be seen on the U.S.G.S. topographic map
from 1892 occupying the central part of the delta (Figure 4).

Please exercise caution in the pit. We are visiting the pits on a Saturday so there should be no
track traffic however please do not approach or climb on the steep loose faces. We will visit two locations
in the southern exposures in the extensive pit to see glaciofluvial topset beds in the upper pit and topset and
forest exposures in the lower part of the south pit. We will exit the main pit and cross Mayall Road to
examine foreset and bottomset beds exposed in the north pit.

Figure 4. Southeastern sector of the U.S.G.S. Gray 15 minute Quadrangle published in 1892. Contour interval is 20
feet. East Gray delta is the flat topped feature surrounding a deep kettle hole.

After exiting the north pit, turn right on Mayall Road and follow to Rte 100

30.6 Continue on Mayall Road by crossing Rte 100. You are traveling across seafloor deposits (31.0 miles) and
will go up the frontal slope of the delta onto the delta braidplain surface (31.3 miles) Note the excellent
view across the delta surface to the left.
31.8 Turn right onto Trout Run Road. Follow to the bottom of the hill, turn left on Fish Hatchery Road and follow to the fish hatchery building on the left (32.2 miles). We will take our lunch break at the Fish Hatchery. Park next to the building or along the road.

**STOP 4: LUNCH STOP, NEW GLOUCESTER FISH HATCHERY, MAINE DEPARTMENT OF INLAND FISHERIES and WILDLIFE.** Gray 7.5 minute quadrangle, 43°56'15.40"N, 70°19'38.44"W

The hatchery is situated in a deep and narrow Eddy Brook valley downcut into the Sabbathday Pond Delta. The spring-fed pools used in the facility are situated along the floor of the erosional valley. Brown trout and rainbow trout are raised in the pools for stocking in Maine streams, ponds and lakes. Take a stroll around the complex to visit the trout pools and the exposure of deltaic sediment above the northernmost trout pool.

Figure 5. Section of the Gray 7.5 minute quadrangle. Lunch stop is located at the Maine IFW Fish Hatchery alongside Eddy Brook which is deeply incised into the Sabbathday Pond Delta.
Return to the vehicles and retrace your path out of the hatchery along Fish Hatchery Road to Trout Run and Mayall Road.

32.6 Turn right on Mayall Road.
32.7 Pull off the road to the right.

**STOP 5: SABBATHDAY POND DELTA**

43°56′7.17″N, 70°19′43.37″W

This is a brief roadside stop on the top of the Sabbathday Pond Delta. Our view is to the west across the former braidplain of the delta surface that build to the south-southeast into the sea. Crossen (1984) and Weddle (1997) describe this extensive landform as a compound delta comprised of the Crystal lake Delta, the Sabbathday Pond Delta and the New Gloucester Delta, all fed from ice contact deposits from the west and north. Thompson et al. (1989) report the topset/foreset contact in this delta at 312 feet (95 m). The LIDAR image (Figure 5) reveals distinct channels that trend from the ice-contact zone in the west to the south and east. The broad channels are slightly lower than the main delta surface and indicated that sea level was falling (rebound had started) while the delta was still actively being fed from the ice margin in the valley to the northwest.

![LIDAR image of the Crystal Lake-Sabbathday Pond Compound Delta, New Gloucester and adjacent terrain.](image)

Continue ahead on Mayall Road
33.3 Cross over Maine Turnpike over the Maine Turnpike
34.7 Turn right on Sabbathday Road
35.3 Sabbathday Pond on right.
35.8 Turn right at intersection with Rte. 26. And continue ahead over the crest of the drumlin
37.3 Turn right on Outlet Road near the base of the drumlin. Follow Outlet Road downhill using caution on the narrow camp road.
38.3 Turn left immediately before the bridge over the pond outlet. Follow the access road into the Maschino Pit.
STOP 6: SABBATHDAY POND ICE CONTACT DEPOSITS
Duayne Maschino and Sons Gravel 43°59′19.5″N, 70°20′56.0″W

The ice contact sand and gravel deposits represent fluvial sediments that fed the extensive Sabbathday Pond Delta to the south. The landforms in this area north of the pond include eskers, kames and kame terrace deposits (Figure 7).

Exit the pit and return to Outlet Road and take a left turn. Follow ahead to the stop sign
38.7  Turn left on Snows Hill Road (aka Shaker Road). Follow road over the Maine Turnpike (40.4 miles).
41.2  Turn right at the intersection of Bald Hill Road turn right. Bear right near the top of the hill to the junction
41.6  Junction of Route 100 of Rte. Cross Rte. 100 straight ahead onto Rte. 231.
     Follow 231 through New Gloucester village (43.4 miles) and downhill to the lowlands.
44.7  Cross railroad tracks and turn left into the Intervale Road Pit.

Figure 7, Northwestern sector of the U.S.G.S. Gray 7.5 minute quadrangle. Stop 6 is located in ice contact stratified drift exposed in the Maschino and Son Gravel pits northeast of the pond.
Stop 7: INTERVALE ROAD GLACIALMARINE FAN
Gray 7.5 minute Quadrangle, 43°57’6.4”N; 70°15’40.7”W  125 feet (40 m) asl.

The landform at this stop was mapped by Weddle (1995) as a glacialmarine fan deposit in the east central sector of Gray 7.5 minute quadrangle, New Gloucester. The landform is situated in the lowland well below marine limit. Coarse material deposition in the fan was probably short lived and the landform was draped by distal fine-grained Presumpscot Formation. Marine shell fragments can be found in the deposit.

Figure 8. East central sector of the Surficial Geologic Map of the Gray 7.5 minute quadrangle (Weddle, 1997) Stop 7 is located in landform mapped as Pmf1.

End of formal field trip. Return to Dick’s Sporting Goods in Topsham.
ALTERNATE STOP 8 EOLIAN DUNE ON SURFACE OF BRUNSWICK SAND PLAIN
Dick’s Sporting Goods Parking Lot, 43°55’59.18”N; 69°58’58.84”W.

If time (and remaining light) allows we will close up at this stop at a small exposure in an eolian dune situated at the apex of the Brunswick Sand Plain, north of the Androscoggin River. Early postglacial eolian deposits are common in this area of southern Maine. In general the eolian deposits are comprised of loose fine-grained sand. Eolian deposits are sometimes referred to as an “eolian mantle” that blanket late Pleistocene deposits and landforms. In other cases dunes were formed from deflation of proglacial sediments in the unvegetated early postglacial time. In nearby Freeport the “Desert of Maine” is a well-known tourist destination where the loose sand is exposed in massive dune forms at the surface.

Since exposures are limited to non-existent in the Brunswick Sand Plain itself we will also take the opportunity to discuss the significance of this major regressive feature.

Figure 8. LIDAR image of the lower Androscoggin Valley area featuring the Brunswick Sand Plain.
REFERENCES CITED


Thompson, W. B., and Borns, H. W., Jr., 1985, Surficial geologic map of Maine: Augusta, Maine Geological Survey, 1:500,000 map.