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B2: Glacial Deposits and Late-Glacial to PostGlacial Alluvial Fans in the Northwestern White Mountains, New Hampshire

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GLACIAL DEPOSITS AND LATE-GLACIAL TO POSTGLACIAL ALLUVIAL FANS IN THE NORTHWESTERN WHITE MOUNTAINS, NEW HAMPSHIRE

By

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INTRODUCTION

This trip visits the northwestern White Mountains, where a wide variety of glacial and glacial-lake deposits formed during recession of the Laurentide Ice Sheet (Thompson *et al.*, 2017). We will examine key sites that turned up during recent field studies and see examples of new LiDAR imagery that is being used to refine the mapping of glacial features in the area. The results of detailed surficial quadrangle mapping for the New Hampshire Geological Survey's STATEMAP program have likewise contributed much to our understanding of the area's glacial history (*e.g.* Hildreth, 2009; Fowler and Barker, 2015; Thompson, 2016).

The trip ranges from Randolph west to Bethlehem, and north to Jefferson and Lancaster (Figs. 1, 2). Topics of discussion include ice-marginal deposits of the White Mountain Moraine System, their age and climatic significance, glacial lake sequences and drainage channels, alluvial fans, and a saprolite occurrence.

Note that all gravel pit stops are on private property, and permission must be obtained from the owners for any future visits.

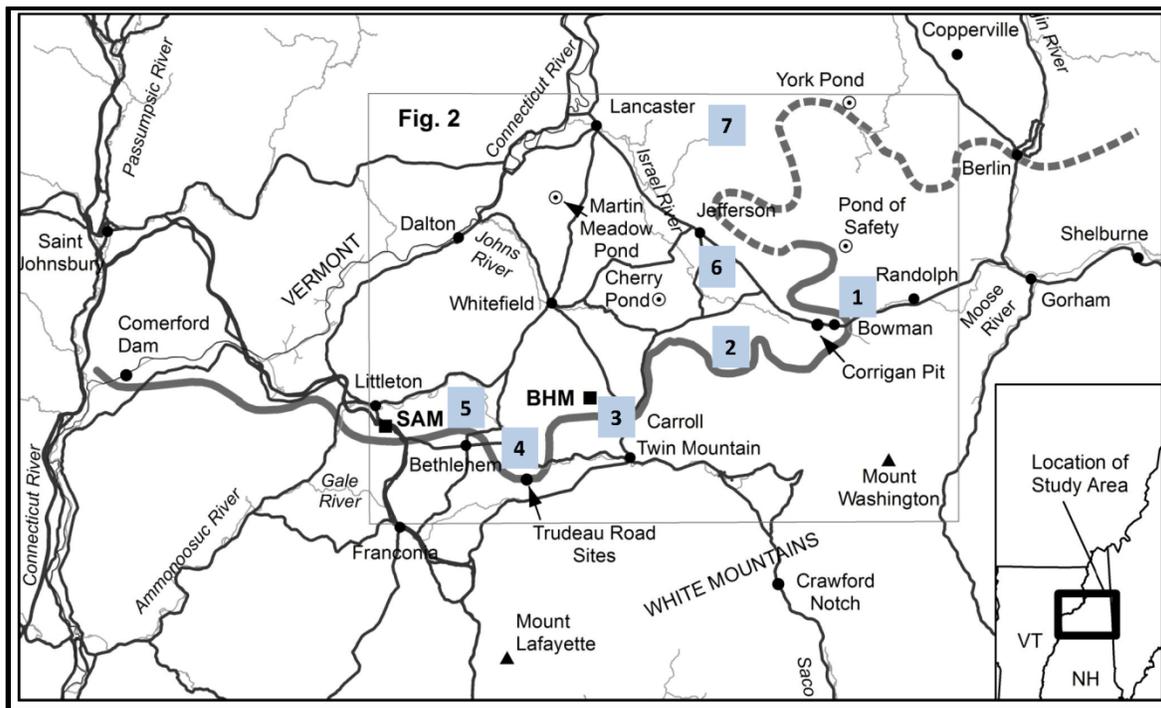


Figure 1. Map of the northern White Mountains showing locations of Trip B2 stops 1–7. Modified from Thompson *et al.* (2017). Long solid gray line indicates limit of Littleton-Bethlehem (L-B) readvance and related moraines of the White Mountain Moraine System (WMMS). Dashed line shows correlation with the Berlin moraines. Open circles mark ponds cored by Christopher Dorion. Black squares indicate sample locations for cosmogenic-nuclide exposure dating by Greg Balco: Sleeping Astronomer Moraine (SAM) and Beech Hill moraines (BHM). Rectangle shows area of geologic map (Fig. 2).

There are three principal rivers in the field trip area. The Ammonoosuc River originates on the side of Mount Washington in the Presidential Range. It flows west to Littleton and then southwest to the Connecticut River. Neighboring rivers to the north – the Johns River and Israel River – flow northwest and join the Connecticut River at Dalton and Lancaster respectively. The valleys of all these rivers hosted ice-dammed glacial lakes during recession of the Laurentide Ice Sheet. They also were affected by a major late-glacial readvance of the ice sheet, during which an extensive series of moraines were deposited across northern New Hampshire. The stops on our trip were chosen to present stratigraphic and geomorphic evidence of these glacial events, as well as the modification of the landscape by postglacial streams.

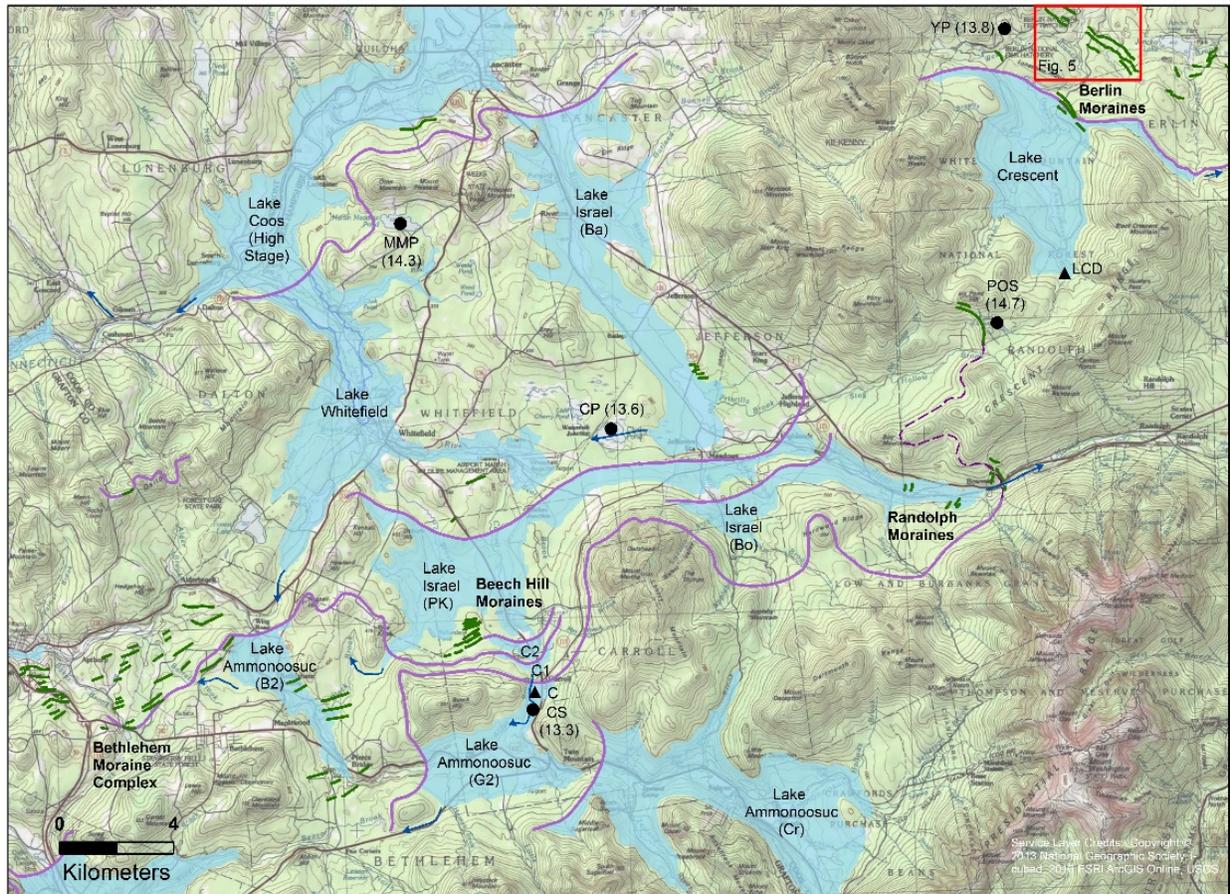


Figure 2. Geologic map of the field trip area showing moraine clusters (green lines) deposited during the L-B readvance and recession from the readvance maximum, associated glacial lakes (blue), lake spillways (blue arrows), recessional ice-margin positions (purple lines), and area covered by LiDAR image in Fig. 5 of Thompson *et al.* (2017). Lake stages shown here include: Crawford (Cr), Gale River 2 (G2), and Bethlehem 2 (B2) stages of Lake Ammonoosuc; Lake Carroll stages C1, C2; and the Bowman (Bo), Pine Knob (PK), and Baileys (Ba) stages of Lake Israel. Black dots mark coring sites and approximate basal radiocarbon ages in cal ka BP for Pond of Safety (POS), Carroll spillway (CS), Cherry Pond (CP), Martin Meadow Pond (MMP), and York Pond (YP). Black triangles are locations of the Carroll delta (C) and Lake Crescent delta (LCD). Unlabeled glacial lakes include small parts of Lake Franconia (southwest corner of map) and two arms of Lake Hitchcock on the western border. Topographic base map contour interval is 20 m. Modified from Thompson *et al.* (2017).

PREVIOUS WORK

Many geologists have investigated the glacial history of the White Mountains, starting in the 1800s. The following account summarizes and updates reviews by Thompson (1999) and Thompson *et al.* (2009a). Visits by Louis Agassiz, Edward Hitchcock, and Charles Lyell stimulated interest in glaciation of the region. Early investigations in the present field trip area focused on clusters of drift ridges and hummocks in the Ammonoosuc River valley that eventually became known as the Bethlehem Moraine. Agassiz published the first observations on these deposits following his 1847 visit to the White Mountains. He thought that the morphology and boulder provenance of moraines in the Bethlehem area proved they were deposited by a local glacier flowing north from the vicinity of Mount Lafayette in Franconia.

Charles Hitchcock (1878a) agreed with Agassiz's theory of local ice depositing moraines from the south. Upham likewise concurred with this theory and formally named the Bethlehem Moraine (Upham, 1904, p. 12). His description of the Bethlehem Moraine is very similar to what recent workers have observed: "The material of this belt is chiefly till, with some modified drift, as kames, or knolls of gravel and sand. The contour is very irregular, in multitudes of hillocks and little ridges, grouped without order or much parallelism of their trends. Everywhere in and upon these deposits boulders abound, ...being far more plentiful than in and on the adjoining smoother tracts of till throughout this region" (Upham, 1904, p. 11–12).

James Goldthwait (1916) reinterpreted the Bethlehem Moraine. Agassiz's model was found to be flawed because it proposed a topographically unrealistic ice-flow path, lacked adequate documentation of northward erratic transport, and was not supported by striation evidence. Goldthwait pointed to the lack of recessional moraines in the Franconia Range as another problem with Agassiz's and Upham's local-ice models. He said that the geometry and provenance of the Bethlehem Moraine favor deposition from the north by the continental ice sheet. Goldthwait's 1916 paper included the first map of the Bethlehem Moraine.

Ernst Antevs (1922) inferred from his Connecticut Valley varve records that a glacial readvance occurred west of Littleton where Comerford Dam is now located (Fig. 1). Crosby (1934) reached the same conclusion, supported by the two-till stratigraphy that he found at the dam site. Antevs and Crosby equated this readvance with that which deposited the Bethlehem Moraine complex. Lougee (1935) referred to this event as the "readvance at Littleton", and Thompson *et al.* (1999) named it the "Littleton-Bethlehem readvance" to stress the connection with the deposits historically known as the Bethlehem Moraine. Richard Lougee (1935) described a new section next to Comerford Dam showing deformed varves between two till units. He correlated the Comerford varves with Antevs' nearby sections and inferred from a gap in the varve sequence that the readvance covered the site for 151 years.

In 1930 Lougee assisted J. W. Goldthwait in a gravel inventory funded by the New Hampshire Highway Department. He was assigned to map several 15-minute quadrangles in the White Mountains. Lougee (n.d.) prepared a manuscript that included a wealth of new information on glacial deposits, ice-dammed lakes, and meltwater drainage routes in the region. A copy of this paper resides in the Special Collections of the Dartmouth College Library. It is unfortunate that it was never published, since it contains the first analysis of the stages of glacial Lake Ammonoosuc and corresponding spillways.

Following a research hiatus of several decades, the 1996 NEIGC trip to the study area presented results of renewed work on deglaciation of the northwestern White Mountains (Thompson *et al.*, 1996). In the guidebook for that trip, Christopher Dorion included his analysis of sediment cores from Surplus Pond in western Maine and Pond of Safety in Randolph, New Hampshire, showing clear stratigraphic evidence of Younger Dryas cooling. Jack Ridge and his students led a companion trip presenting new results from the glacial Lake Hitchcock varve sequence in the vicinity of Comerford Dam (Ridge *et al.*, 1996). During this same trip, Thompson described a stream bluff exposure near the New Hampshire end of the dam that showed further stratigraphic evidence of glacial readvance.

Thompson *et al.* (1999) discovered moraine clusters in Carroll and Randolph, which correlate with the Bethlehem moraine complex, and subsequently found another moraine series of similar age in the Berlin area (Fig. 2). The overall moraine belt spans northern New Hampshire and is now called the White Mountain Moraine System (Thompson *et al.*, 2009a, 2017). Dorion's radiocarbon ages from basal pond sediments in the region suggested an age of 12,000 radiocarbon years (~14 cal ka) for the Littleton-Bethlehem (L-B) readvance and associated moraines, coinciding with the brief interval of cold climate called the Older Dryas Chronozone. Ridge pinpointed the

readvance age much more precisely and confirmed the Older Dryas connection through his work on the New England Varve Chronology and its relation to the Comerford Dam readvance site. His varve correlations in the Connecticut River valley west of Littleton showed that the readvance occurred at 11,900–11,800 ^{14}C ka BP (13.9–13.8 cal ka BP) (Ridge *et al.*, 1999, 2004). The Lake Hitchcock varve chronology is now incorporated in the North American Varve Chronology of Ridge *et al.* (2012). Ridge's latest refinement of the L-B readvance age places it at ~14.0–13.8 cal ka BP, and has established a firm connection to the Older Dryas event (GI-1d) in the Greenland ice core record (Thompson *et al.*, 2017). Greg Balco used this readvance age to help calibrate cosmogenic-nuclide production rates for exposure dating in New England. He discussed this work and its application to the White Mountain moraines during the 2009 NEIGC (Thompson *et al.*, 2009a) and in the paper by Thompson *et al.*, 2017. Balco's study sites are indicated on Figure 1.

MORaine CLUSTERS AND ICE-DAMMED GLACIAL LAKES

Moraines

The following information is from Thompson *et al.* (2017). Most of the moraines in the field trip area are concentrated in three clusters that constitute large parts of the White Mountain Moraine System (WMMS). From west to east, these are the Bethlehem moraine complex, the Beech Hill moraines, and the Randolph moraines (Fig. 2). (The Berlin moraines are also briefly described here for sake of completeness.) The moraines in all these areas are generally similar to one another. They are composed predominantly of loose sandy till with abundant stones including many granitic boulders derived from local plutons. The moraine ridges typically are 3–30 m high and rarely as much as 50 m. Individual segments are up to 1300 m long and most are sharp crested. The spacing between moraines varies, having depended on ice retreat rate and sediment supply. In the tight cluster of the Beech Hill moraines, it ranges from 30 m to about 200 m.

Bethlehem moraine complex. The Bethlehem moraines (Fig. 2) were deposited during ice recession from the upper Ammonoosuc valley. This is the most diffuse group of moraines in the WMMS, spanning up to 7 km of northward to northwestward ice-margin retreat. A few outlying moraines of the Bethlehem complex occur in the western part of Littleton, 12 km east of the Comerford Dam readvance site. Exposures and well logs indicate the Bethlehem moraines consist chiefly of till. Thompson *et al.* (1999) described a moraine cross-section in Littleton village where shear structures indicated ice shove from the north. The high (30–40 m) sharp-crested moraines just southeast of Littleton were deposited in terrestrial settings at the terminus of the moraine belt, whereas some of the other moraines in the Bethlehem moraine complex formed where the ice margin stood in glacial Lake Ammonoosuc.

New information on the stratigraphy of the Bethlehem moraines has come from test borings by Sanborn, Head, & Associates (2014) at the regional landfill on Trudeau Road in Bethlehem, about 30 km east of Comerford Dam (Fig. 1). The landfill is situated on one of the most distal moraines in this part of the complex. Three of the deepest borings at the Trudeau Road site are located in the proximal part of the moraine, along an E-W line over a distance of ~340 m. The western boring (B-916D) penetrated 31 m of surficial sediments overlying bedrock. It encountered 13 m of readvance till overlying 10 m of sand, silt, and minor gravel. The latter unit in turn overlies 8 m of till. The same units occur to the east, where boring B-918D ended at ~38.5 m without reaching bedrock (3.7 m till / 29.3 m silt-sand / 5.5 m till). Still farther east, boring B-919D encountered 4.5 m of till / 42.1 m of silt-sand / 3.0 m of till / 2.2 m of silt-sand.

Two sand and gravel pits near the Trudeau Road landfill show readvance till overlying water-laid glacial sediments. The section in one pit exposes a moraine consisting of till overlying coarse gravel with sand lenses. We will visit this site at Stop 4. The other section is 1.25 km southeast of the first, in the distal flank of the moraine on which the landfill is located. It shows silty-sandy diamict with small rounded stones overlying well-stratified glaciofluvial sand and fine gravel. The diamict is interpreted as till derived from recycling of glaciolacustrine sediments during local ice readvance. The landfill borings and nearby pit exposures collectively show that at least some of the distal Bethlehem moraines are not composed solely of till. The upper (readvance) till at all of these Bethlehem sites is thought to be equivalent in age to the upper till at Comerford Dam.

Beech Hill moraines. Thompson *et al.* (1999) discovered a cluster of end moraines just north of Beech Hill in the town of Carroll (Fig. 2). These moraines are located about 7 km east of the Bethlehem moraine complex. The Beech Hill moraine cluster is at least 0.8 km wide from its distal to proximal margin. New LiDAR imagery shows

additional moraine ridges and hummocks in the wooded area just to the north, which are currently being field-checked. Elevations of the well-defined moraines are 396–427 m. They comprise ten till ridges that are 4–12 m high, up to 700 m long, and trend ENE-WSW. Granite boulders up to 3 m wide are abundant on the surfaces of the moraines. Considering the location of the Beech Hill moraines and their relationships to local glacial-lake stages, it is reasonably certain that they correlate with the northern part of the Bethlehem moraine complex.

Randolph moraines. Another cluster of moraines, which we correlate with the Bethlehem and Beech Hill moraines, occurs to the northeast in the Israel River valley in Randolph (Fig. 2). The Randolph moraines span about 3.4 km in the direction of downvalley ice retreat (toward the west-northwest). The largest and most distal of these moraines are located on the drainage divide at the head of the Israel valley, in the Bowman area of Randolph, where their crests reach elevations of about 540 m (see Stop 1). A major outlying moraine near the Pond of Safety coring site in the hills of northern Randolph (Figs. 1, 2) rises to 720 m. Its location relative to the moraines at the head of the Israel valley suggests the moraines in both places were deposited by the same ice tongue and perhaps at the same time. The difference in their elevations may reflect a steep ice-surface profile near the glacier margin during the L-B readvance. These distal moraines are up to 50 m high and associated with proglacial channels that fed meltwater into neighboring river basins to the north and east.

Exposures in Randolph have recorded a glacial readvance of at least several hundred meters in the upper Israel valley. The Corrigan pit, located on the south side of Valley Road and 0.8 km east of the Jefferson-Randolph town line (Fig. 1), shows glaciolacustrine deltaic sand and gravel overlain by 6 m of stony glacial till comprising a moraine ridge. Recumbent folds and thrust faults in the upper part of the lacustrine unit indicate an ice readvance from the west, concurrent with deposition of the moraine.

Berlin moraines. A series of moraines near the town of Berlin in the northeastern White Mountains have been mapped with the aid of LiDAR imagery (Fig. 2; Thompson et al., 2009b; Thompson and Svendsen, 2015). The most clearly defined part of the moraine cluster is 3.5 km wide. Moraine crest elevations range from up to 492 m south of Jericho Lake to about 415 m in the Upper Ammonoosuc River valley. The Berlin moraines are low but very distinct till ridges that mostly trend northwest-southeast. They were deposited by ice retreating northeast from the headward part of the Upper Ammonoosuc basin. The moraines are 3–10 m high and strewn with large granitic boulders.

Hildreth's mapping (2009) showed a prominent moraine just west of the Pond of Safety coring site in Randolph (Fig. 2), on the divide between the Israel River and Upper Ammonoosuc River basins. As noted above, this deposit is an outlying member of the Randolph moraines. LiDAR imagery reveals multiple channels that drained meltwater north from the Pond of Safety moraine and supplied sediment to the glacial Lake Crescent delta described below. The position and elevation of this delta require that the Upper Ammonoosuc valley was dammed by the ice margin that deposited the Berlin moraines, thus showing that the Berlin moraines are coeval with the Randolph moraines and likewise part of the WMMS (Fig. 1).

Glacial lakes

Glacial Lake Ammonoosuc. Warren Upham's work with the Hitchcock survey led him to propose that a lake had existed in the Fabyan area in the upper part of the Ammonoosuc River valley (Upham, 1878). Goldthwait (1916) named this water body "Lake Ammonoosuc". It resulted from damming of the west-draining valley by a tongue of late Wisconsinan ice receding from the Bethlehem area. As the ice margin withdrew, successively lower spillways for the lake were uncovered and the lake level fell.

Goldthwait (1916) identified two levels of glacial Lake Ammonoosuc: a higher level into which a "pitted outwash plain" and other ice-contact deposits at Carroll and Twin Mountain were built, followed by a lower level into which the Bethlehem moraine complex was deposited. Lougee (n.d.) realized that an earlier and higher stage of Lake Ammonoosuc (his "Crawford Stage") drained east through Crawford Notch at an elevation of approximately 573 m (1880 ft). This was the same route followed by a subglacial tunnel drainage which formed the esker in the upper Ammonoosuc Valley (Goldthwait and Mickelson, 1982).

Thompson *et al.* (1999) named the post-Crawford stages of Lake Ammonoosuc. However, it was Lougee's undated manuscript (*ca.* 1930) that first identified some of these lake levels and their outlets. After the Crawford

Stage, glacial Lake Ammonoosuc drained southwestward through five progressively lower spillways (G1-G5 in Figure 3) into the Gale River valley. The spillway for the Gale River 2 Stage is a prominent channel that can be seen along U. S. Route 3 southwest of Twin Mountain village. Later spillways north of Bethlehem village drained the Bethlehem and Wing Road Stages of Lake Ammonoosuc into Indian Brook and later directly into the Ammonoosuc River. The Gale River and younger stages of the lake generally were not deep, so the ice margin probably was grounded on the lake floor. The widest and deepest stage may have been Gale River 2. A well (CFW 53) just west of Twin Mountain encountered a contact between thick glaciolacustrine clay and the underlying till at an elevation of 401 m (Flanagan, 1996). Comparison with the nearby G2 spillway elevation of 445 m (Fig. 3) indicates an initial water depth of at least 44 m.

Figure 3 also shows proposed recessional positions of the late Wisconsin ice margin that were contemporaneous with the Gale River and later stages of Lake Ammonoosuc. Thompson *et al.* (1999) inferred these ice margins from the orientation of segments of the Bethlehem moraine complex, together with ice blockages of the valley that would have been required to hold the lake at elevations corresponding to known deltas and spillways. The agreement between elevations of Lake Ammonoosuc deltas and terraces in Twin Mountain, and the matching spillways of the Gale River 2-4 Stages, suggests that the receding glacier margin was a tight dam for the lowering lake. During the evolution of Lake Ammonoosuc, water carried sediment into the lake not only from the melting glacier but also from the early Ammonoosuc River and smaller streams draining the surrounding mountains, as shown by Lougee (1940). Both the lake and the mouth of the river shifted westward as the lake level dropped, and Lake Ammonoosuc ultimately disappeared when the ice margin receded from the Alderbrook area in northernmost Bethlehem.

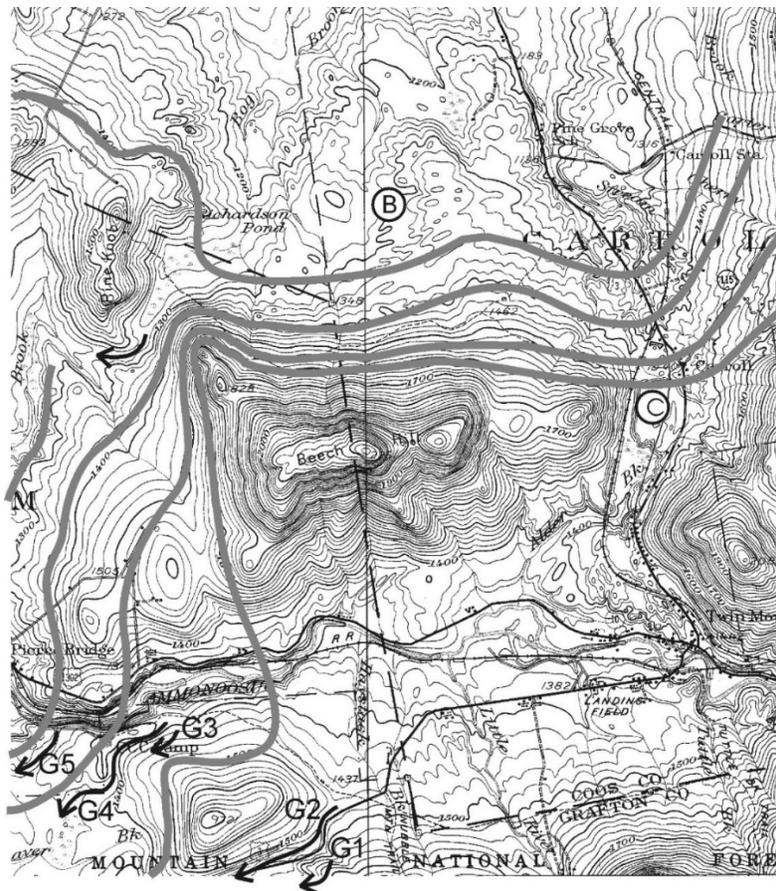


Figure 3. Part of the Whitefield 15-minute quadrangle, showing inferred ice-margin positions (gray lines) and meltwater spillway channels (arrows). “B”: Beech Hill moraines. “C”: Carroll Delta (Stop 3). Labeled arrows (G1 etc.) show spillways for the Gale River stages of glacial Lake Ammonoosuc listed below. Spillway elevations are based on contours from the newer Bethlehem 1:25,000 metric quadrangle. From Thompson *et al.* (1999).

Lake stage:		Elevation:
Gale River 1	G1	477 m (1565 ft)
Gale River 2	G2	445 m (1460 ft)
Gale River 3	G3	435 m (1427 ft)
Gale River 4	G4	423 m (1387 ft)
Gale River 5	G5	405 m (1328 ft)

Lakes Carroll, Israel, and Whitefield. Three closely related glacial lakes developed as the ice receded northward from the Ammonoosuc valley. Lake Carroll formed north of Twin Mountain village, just south of the Beech Hill moraines. The first stage of this lake (C1, Fig. 2) was very small and briefly drained south through the 429-m channel that incises the Carroll delta (Stop 3). Lake Carroll then dropped to the C2 stage as it drained west along the ice margin.

At the same time that glacial Lake Carroll existed, deglaciation of the neighboring Israel valley to the east resulted in the first stage of ice-dammed Lake Israel. This stage (Bo, Fig. 2) drained east across the divide at Bowman and into the Moose River valley. Deltaic and subaqueous fan deposits were built into the Bowman stage lake. With further ice retreat, lake waters in the Israel valley may have briefly expanded west as the Pine Knob (PK) stage and drained through a 387-m col into the Ammonoosuc valley, as shown in Figure 2. Slight additional recession led to the creation of two separate lakes: the Baileys stage of Lake Israel to the east and Lake Whitefield to the west. The Baileys stage (Ba) spilled west across a low divide at Cherry Pond in Jefferson and thence into Lake Whitefield. Lake Israel terminated when ice retreat to Lancaster caused it to merge with glacial Lake Coos in the Connecticut River valley. Lake Coos (Fig. 2) was a large drift-dammed lake just north of glacial Lake Hitchcock. During ice retreat from the study area, it extended from Dalton, New Hampshire, up the Connecticut valley to North Stratford (Thompson *et al.*, 2011). Lake Whitefield occupied the Johns River valley as the ice margin receded north from Whitefield village. This lake initially drained south into the Ammonoosuc River, and with further ice recession likewise dropped and merged with Lake Coos.

Lake Crescent. Recessional ice margin positions indicated by the Berlin moraines suggest that an ice-dammed lake existed in the Upper Ammonoosuc River valley, in the northeast part of the study area (Figs. 1, 2) and distinct from the Ammonoosuc River basin discussed earlier. The headwaters of this river occupy a mountain-rimmed basin that drains to the north and would have been blocked by the NE-retreating ice lobe that deposited the moraines. A series of nine meltwater channels drained the basin along its eastern border. These channels are lower to the north, recording a succession of probable lake spillways that opened as the ice receded (Thompson and Svendsen, 2015). The dense forest cover and poor exposure of Pleistocene deposits in the Upper Ammonoosuc valley hindered recognition of lacustrine deposits when the area was mapped by Hildreth (2009). However, LiDAR imagery shows a delta complex at the head of the basin that was built by glacial meltwater streams flowing northward into the former lake. Elevations of terraces on the delta top correspond with some of the spillway channels mentioned above, confirming the lowering of the lake level with time. Based on this evidence, we have identified an ice-dammed lake called glacial Lake Crescent after the neighboring Crescent Range (Fig. 2).

ICE-FLOW DIRECTIONS

There are virtually no bedrock exposures within the Bethlehem moraine complex, but striated outcrops in surrounding areas of Bethlehem and Littleton indicate ice-flow directions mostly in the range of 170–190° (Hitchcock, 1878b, 1905; J. W. Goldthwait, 1916; R. P. Goldthwait *et al.*, 1951). J. W. Goldthwait's map shows a few representative examples of the southward flow. Thompson *et al.* (1999) recorded striation trends in this area ranging from 125° to 203°. In the few places where the latter authors could determine relative ages of multiple striation sets, the striations trending S to SSW are usually youngest.

Just north of Bethlehem, Thompson *et al.* (1999) found striations trending 174–185° on the north side of the Ammonoosuc River, and 178–179° on the crest of Dalton Mountain. These data show that late-glacial southward ice flow crossed both the high hills and valley floor north of the Bethlehem moraine complex. Striations between Littleton and the Connecticut River likewise indicate generally southward ice flow. Most striations seen in the latter area trend between 170° and 190°, though a few are more southeasterly. However, south-trending striations also

occur in many places farther south in the Connecticut River basin (Goldthwait *et al.*, 1951), so they are not associated exclusively with the Bethlehem moraine complex.

At 10 localities in the Whitefield-Lancaster-Jefferson area, Thompson *et al.* (2009a) recorded striation sets trending SE to SSE (135–172°). One of these sites clearly shows three successive ice-flow directions. It is located on the NW end of a long road cut on the SW side of U. S. Route 2, 2.6 km SE of Lancaster. This ledge has a protected lee surface with the oldest striation set trending 241° and a younger 165° set. The youngest of the three sets trends 155° on the stoss surface of the outcrop. Recent mapping of the Jefferson quadrangle documented striation sets indicating SE ice flow at multiple sites on the mountainside NE of Jefferson village, and ESE ice flow on slopes SW of Pliny Mountain (Thompson, 2016).

ACKNOWLEDGEMENTS

We are especially grateful to the following persons and companies who granted permission to visit the stops on this field trip and enabled us to carry out research visits on other occasions: Jim Brianas, John Scarinza (Randolph Town Forest), and Judith Asetta (Stop 1); Tim Bradstreet of Pike Industries, Inc. (Stop 3); Mark Champagne of Bethlehem Earth Materials (Stop 4); Nigel Manley of the Society for Protection of New Hampshire Forests in Bethlehem (Stop 5); Craig Kluckie (Stop 6); and Paul Crane (Stop 7). Many other residents of the White Mountains have kindly assisted us over the years, especially Dennis Field of Lancaster, whose knowledge of the North Woods has been most helpful.

Surficial geologic mapping funded by the New Hampshire Geological Survey – USGS STATEMAP cooperative enabled us to gather much of the new data reported here. The authors are indebted to Brian Fowler for sharing his knowledge of the White Mountains and regional geology during many discussions in the field. We also thank Kristen Svendsen of the New Hampshire Department of Environmental Services for digitizing the maps from which Figures 1 and 2 were extracted. Al Falster at the Maine Mineral and Gem Museum conducted the sediment analysis needed to help explain the clay occurrence at Stop 7-A. Recent bedrock mapping of the Jefferson-Randolph area by Dyk Eusden and his students (Bates College) helped us to identify rock types in the study area. Douglas Rankin (U. S. Geological Survey) likewise assisted W. B. Thompson with understanding the bedrock geology of the Littleton area and sources of glacially transported rocks.

ROAD LOG

Meeting point: Assemble at 8:00 am on September 30 at the Randolph Fire Station, 0.5 mile from U. S. Route 2 via the Pinkham B Road (aka Dolly Copp Rd.) (319571 m E; 4915340 m N). The turnoff onto this road is about 28 miles / 40 minutes west of Bethel, ME; and from Gorham, NH, it is 4.5 miles / 7 minutes west of where you turn at the traffic light on the west end of downtown. The trip will start promptly at 8:30 am. Carpooling at the fire station is recommended, and moderate to high clearance vehicles are best for the woods roads. Come prepared with your lunch, water, and clothing suitable for predicted weather. We'll be walking through the woods at Stops 1 and 2 (short to moderate hikes including rocky ground and uneven terrain); all other stops will be accessed by car. The trip will end in late afternoon at a point east of Lancaster, from which participants can drive back to Route 2 in Jefferson and east to NEIGC headquarters in Bethel. Topographic map coverage of the field trip area is provided by the Bethlehem, Jefferson, Mount Dartmouth, and Mount Washington 7.5-minute quadrangles, and the Mount Washington 1:100,000 map. The latter map gives an excellent overview of the White Mountain region.

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 pits and woods roads, but the distances between stops are generally accurate.*

- 0.0** Exit parking lot and turn L on Pinkham B Road.
- 0.5** Turn L on U. S. Route 2.
- 3.4** Just beyond Lowe's Store, turn R on short connector to Durand Road.
- 3.45** Turn L on Durand Road, then go W about 0.45 mile to end of road and park.

STOP 1. BOWMAN MORAINES (Randolph).

At this stop we will examine moraines on the divide between the east-flowing Moose River and west-flowing Israel River. These are the earliest of the Randolph Moraines described above. They are interpreted to have formed at the distal limit of the White Mountain Moraine System during Older Dryas climate cooling and glacial readvance *ca.* 14,000 years ago.

Leaving the parking area, we will see meltwater channels that drained the first stage of glacial Lake Israel. As we go up the driveway to the Brianas residence, we'll pass a couple of small NW-trending moraines that are crossed by the pipeline, and then will climb onto one of the prominent moraines seen from Lowe's Store (Fig. 4). This ridge has a slightly arcuate shape on the topo map, perhaps because its E side has been partly eroded by meltwater flowing off the ice margin. The moraine is up to ~30 m high and trends between NW (in the southern part) and NNW. There are many 1–3 m boulders along its crest, suggesting the ridge is composed of till. The proximal (W) side of the ridge is locally very bouldery and appears more so than the E side. The boulders are white to pinkish-gray, medium-coarse grained, massive to foliated granitic rock with some pegmatite. A deep channel that curves around the N end of this moraine segment was cut by a glacial stream that carried meltwater off to the E. The channel has a prominent boulder lag on its floor. A second large moraine is located SW of the first and is not so high or sharp-crested. Till was formerly exposed in cuts along a logging road on the latter ridge. There are also many quartz boulders on that moraine, which probably came from one of the silicified zones in the local bedrock.

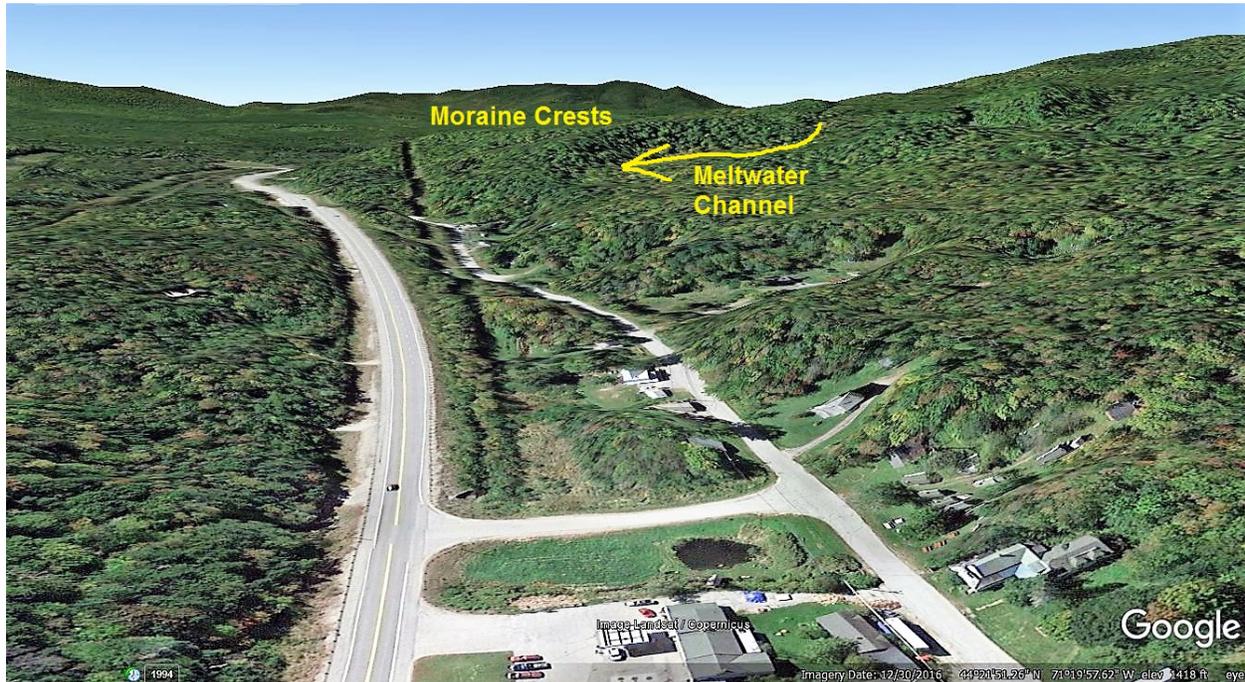


Figure 4. Google Earth view looking west along U.S. Route 2 (left) in Bowman area of Randolph. Lowe's Store is seen in foreground. Labels indicate moraine ridges and a meltwater channel in vicinity of Stop 1.

- 4.4 Return to Route 2 and turn R.
- 8.9 Route 2 crosses one of several hillside meltwater channels (just before large yellow house).
- 9.4 Turn L on Route 115.
- 12.6 Turn L on Cherry Mountain Road.
- 14.7 Turn L on unmarked National Forest dirt road.
- 15.0 Turn around and park at side of road.

STOP 2. MILL BROOK VALLEY DELTA (A BOWMAN STAGE LAKE ISRAEL DELTA) AND ASSOCIATED ALLUVIAL FEATURES (Jefferson).

According to W. Thompson's work in the region, the Mill Brook Valley should contain evidence of the earliest stage of ice-dammed glacial Lake Israel (Thompson *et al.*, 2017). Based upon observation of delta topset/foreset beds in a former pit on Route 2 at the junction with Valley Road (Fig. 5), a glacial lake was present in the eastern portion of the Israel River valley. This lake discharged eastward across a saddle near present day Bowman.

Following deglaciation and resultant emptying of the lake, the whole Israel River system reversed course and drained west as it does today. Fowler and Barker's (2015) mapping of the surficial geology in the Mt. Dartmouth quadrangle documents the Mill Brook Valley Delta, which built northward into Lake Israel. However, it wasn't until LiDAR was available that the full extent of this delta and other features were revealed. The LiDAR will provide the basis for modifying the current features on the Mt. Dartmouth surficial geologic map.

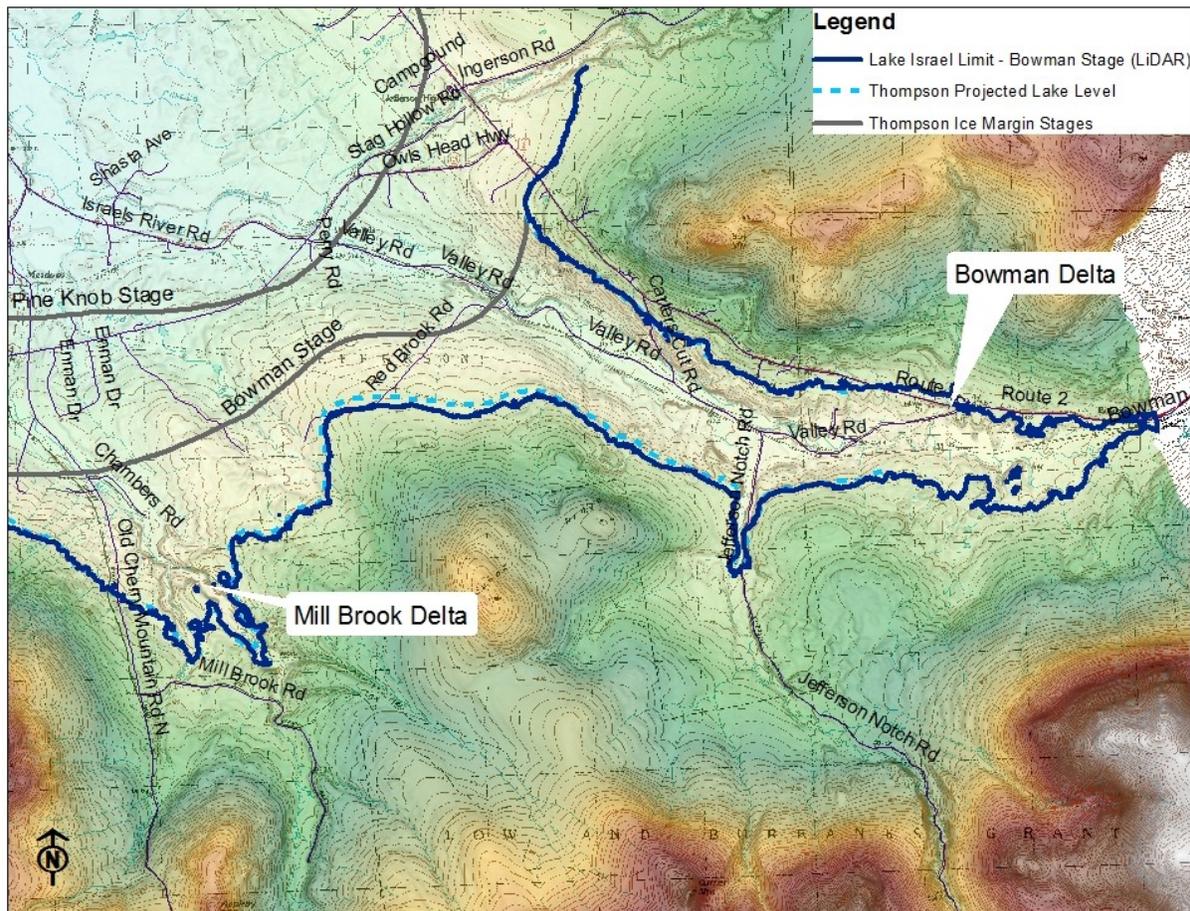


Figure 5. Ice-margin positions and lake levels as presented by Thompson *et al.* (2017) with LiDAR lake level projection by Barker. Base map is a slope map derived from LiDAR Digital Elevation Map (DEM) with an overlay of colorized DEM.

Mill Brook is the primary stream in a northwest-facing watershed which starts at the saddle between Cherry Mountain and Mount Dartmouth. This steep terrain creates a high-energy fluvial system. Much of the watershed is covered by glacial till. Glacial-fluvial deposits are also present, in valleys now occupied by modern streams. Additionally, a large alluvial fan deposit has developed along the western flank of this basin, on the lower part of Streator Brook (Fig. 6). This fan was not delineated in the current mapping but will be part of the revisions to this

map. The road leading to Stop 2 (Cherry Mountain Road) provides ample viewing of Holocene stream terrace deposits. Gentle undulations of former stream channels can be seen in the LiDAR (Fig. 6) but are also apparent in the mowed fields along the road. Looking at Figure 6, you will see an impressive alluvial fan along the southwestern flank of the valley.

From where we stop along Mill Brook Road, we will walk downhill toward the confluence of Mill Brook and Appleby Brook (Fig. 6). As the terrain begins to flatten we are entering a portion of the Mill Brook Delta that built out into Lake Israel (Fig. 7) as sediments were deposited into the Bowman Stage of the lake by the late Pleistocene Mill Brook. The delta was subsequently bisected by present day Mill Brook.

There is very limited exposure of the delta but sampling along traverses up a couple of the terraces showed a basal till overlain by sand and sub-rounded gravel and cobble deposits. Projection of the Bowman spillway elevation (457 meters/1499 feet) using the LiDAR DEM provides excellent agreement with the flat upper surface of this deposit being a submerged delta top. Looking closely at Figures 7 and 8 shows that as the lake receded, the delta top became exposed and Mill Brook began incising channels into it. At least three stream channel positions are shown in the LiDAR. Subsequently, Mill Brook established its present course and cut down through the center of the delta.

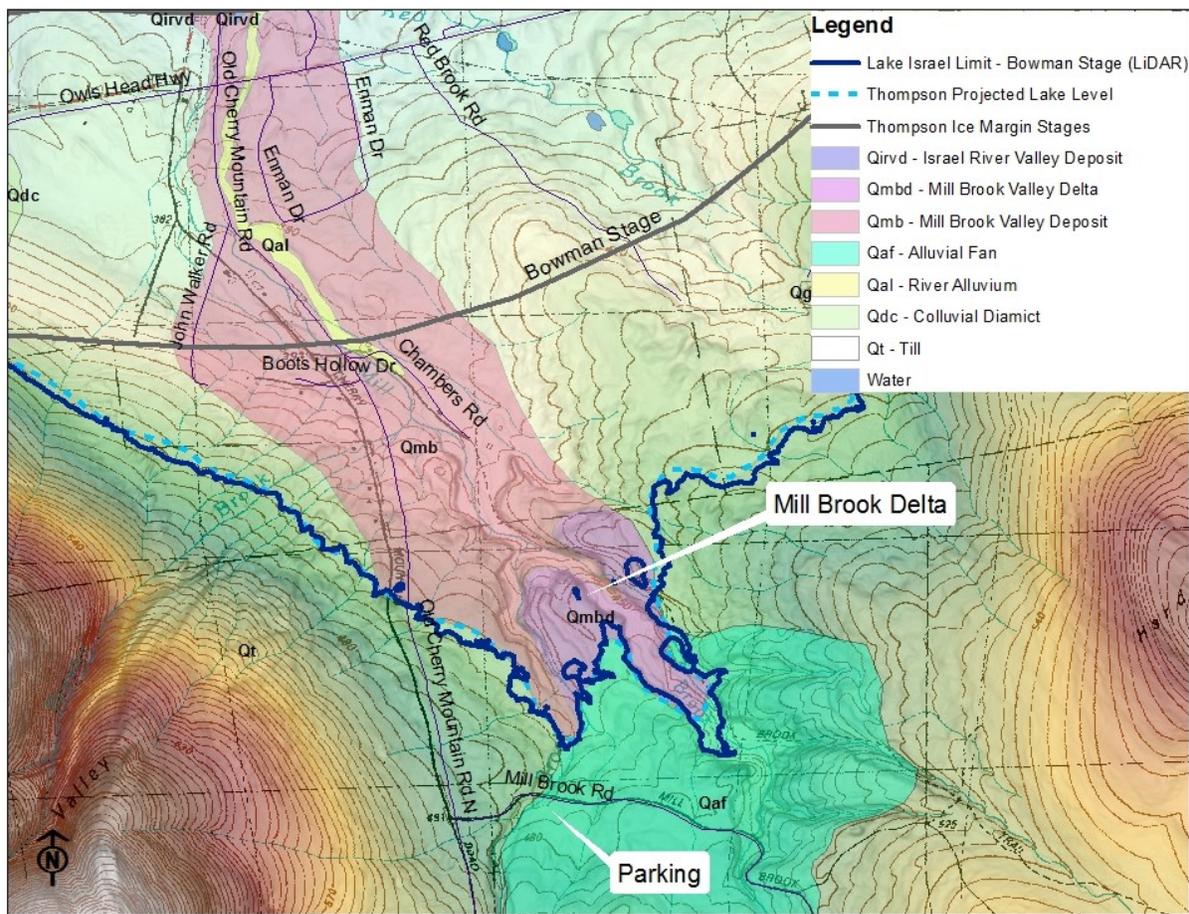


Figure 6. Surficial geologic map of Mill Brook Valley by Fowler and Barker (2015). Base map consists of LiDAR slope map, overlain by colorized DEM and 1:24k-scale topographic map features and contours.

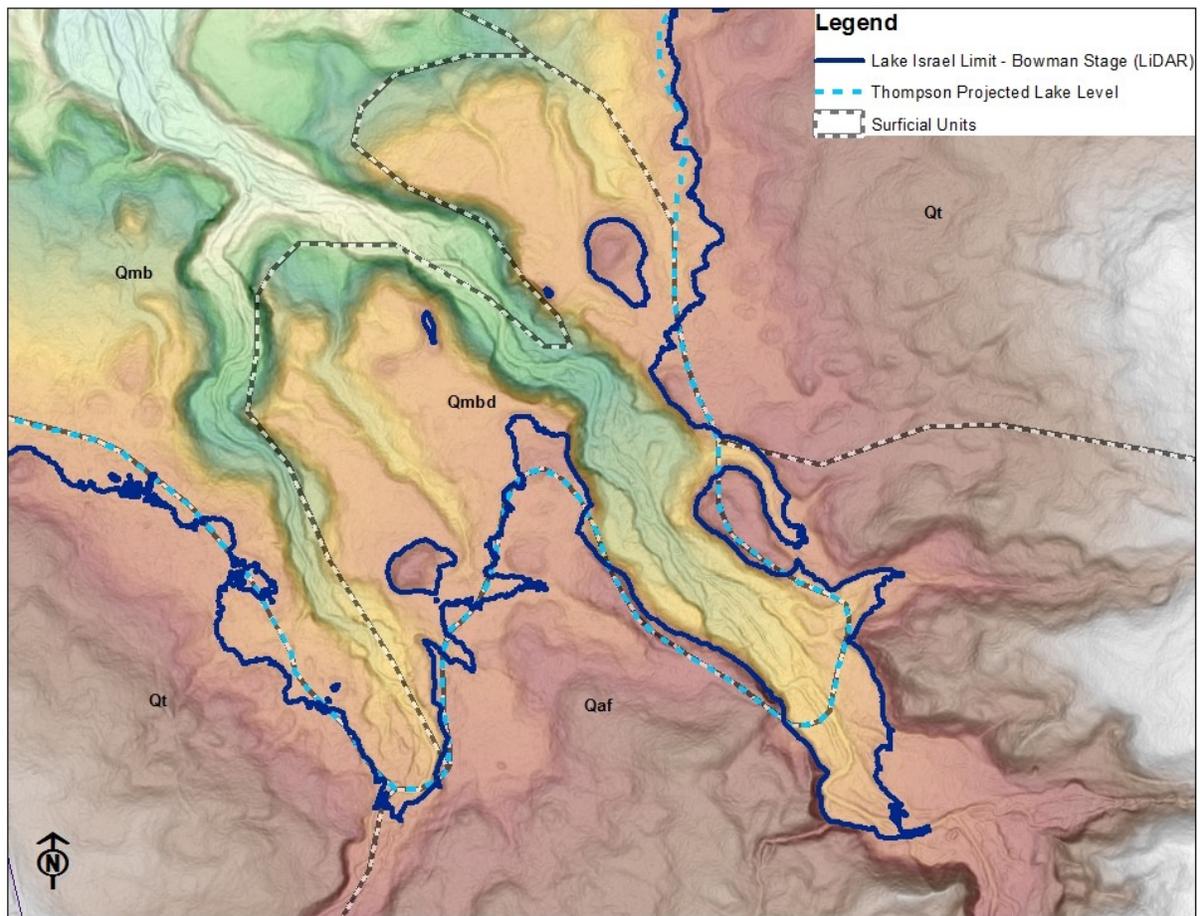


Figure 7. LiDAR slope map and colored DEM overlay map of the Mill Brook Delta and surroundings. Surficial map units are denoted by labels and their outer boundaries to allow full viewing of LiDAR slope map and colored DEM overlay.

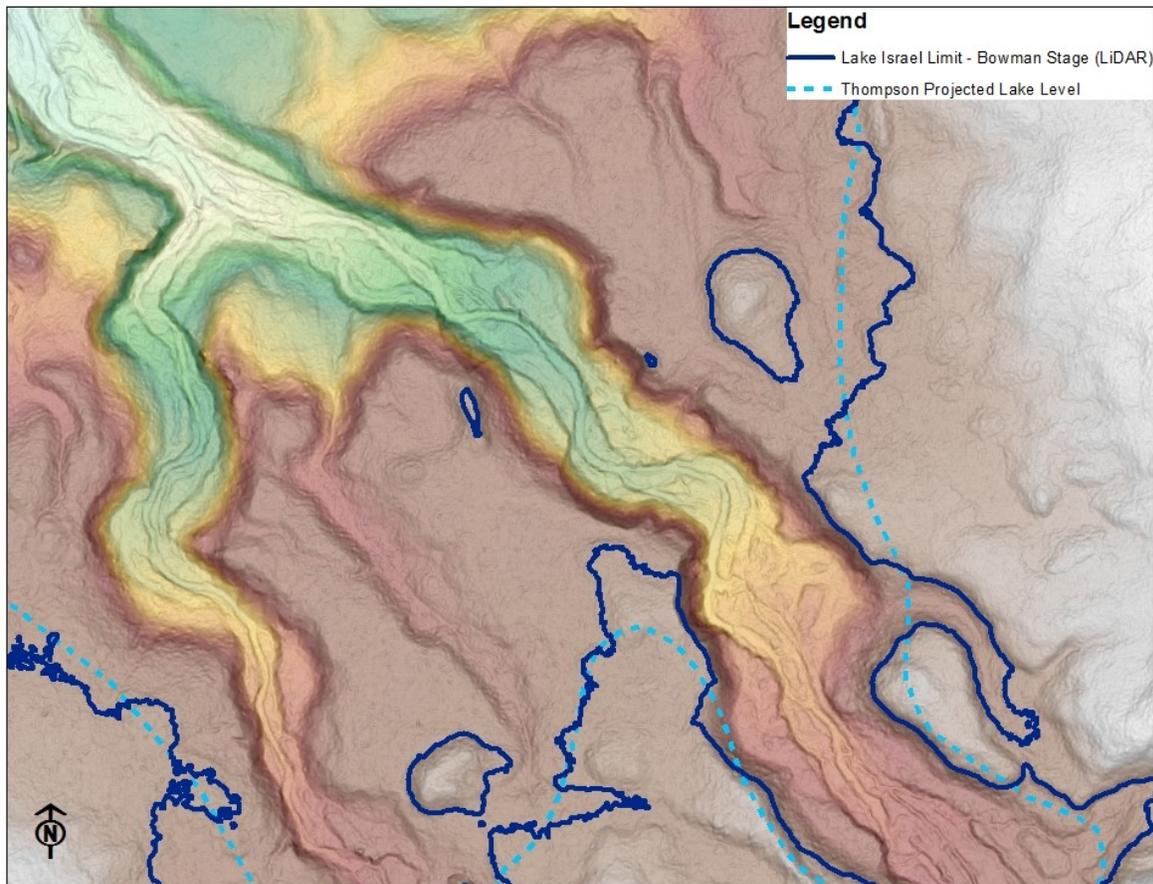


Figure 8. Map of the Mill Brook Delta. Note that the delta top has been dissected by stream channels.

- 15.3 Return to Cherry Mountain Road and turn R.
- 17.3 Turn L on Route 115.
- 23.8 Turn R on U. S. Route 3.
- 23.9 Turn L into Twin Mountain Sand & Gravel pit.

STOP 3. CARROLL DELTA (Carroll).

This stop is in a large ice-contact delta, known as “the Carroll Delta” (Lougee, 1940) which built southward into glacial Lake Ammonoosuc. The delta formed when the receding late Wisconsinan ice margin was pinned against Beech Hill to the west and Cherry Mountain to the east. Figures 2 and 3 show the relationship of the Carroll Delta to inferred ice-margin positions.

Aerial photographs taken in 1955 (prior to the pit operation) show a network of subparallel channels extending from north to south across the full width of the delta plain. These photos also show a steep ice-contact slope that formerly existed on the north edge of the delta. According to Goldthwait (1916), the delta plain (since removed in the pit area) was “strongly pitted” in the proximal part but smooth and non-kettled in the central and distal portions. He also observed that the fluvial gravel on the delta top was less bouldery toward the front of the delta.

The original delta plain has an elevation of approximately 450 m (1475 ft). It was graded to the Gale River 2 stage of glacial Lake Ammonoosuc, which had a spillway at about 445 m (1460 ft) (Fig. 3). This spillway is

prominently visible along U. S. Route 3 southwest of Twin Mountain village, especially when leaves are off the trees. The Carroll Delta received meltwater and sediment from a succession of lateral channels on the hillside to the northeast. These channels are evident on the topographic map but are much clearer on the LiDAR image (Fig. 9). However, it seems unlikely that the volume of sediment removed from the channels would have been sufficient to build the large delta. Most ice-contact deltas in New England were fed mainly by subglacial tunnel drainage. A possible feeder channel for the Carroll Delta is marked by a short esker segment near the railroad track in the woods north of the delta.

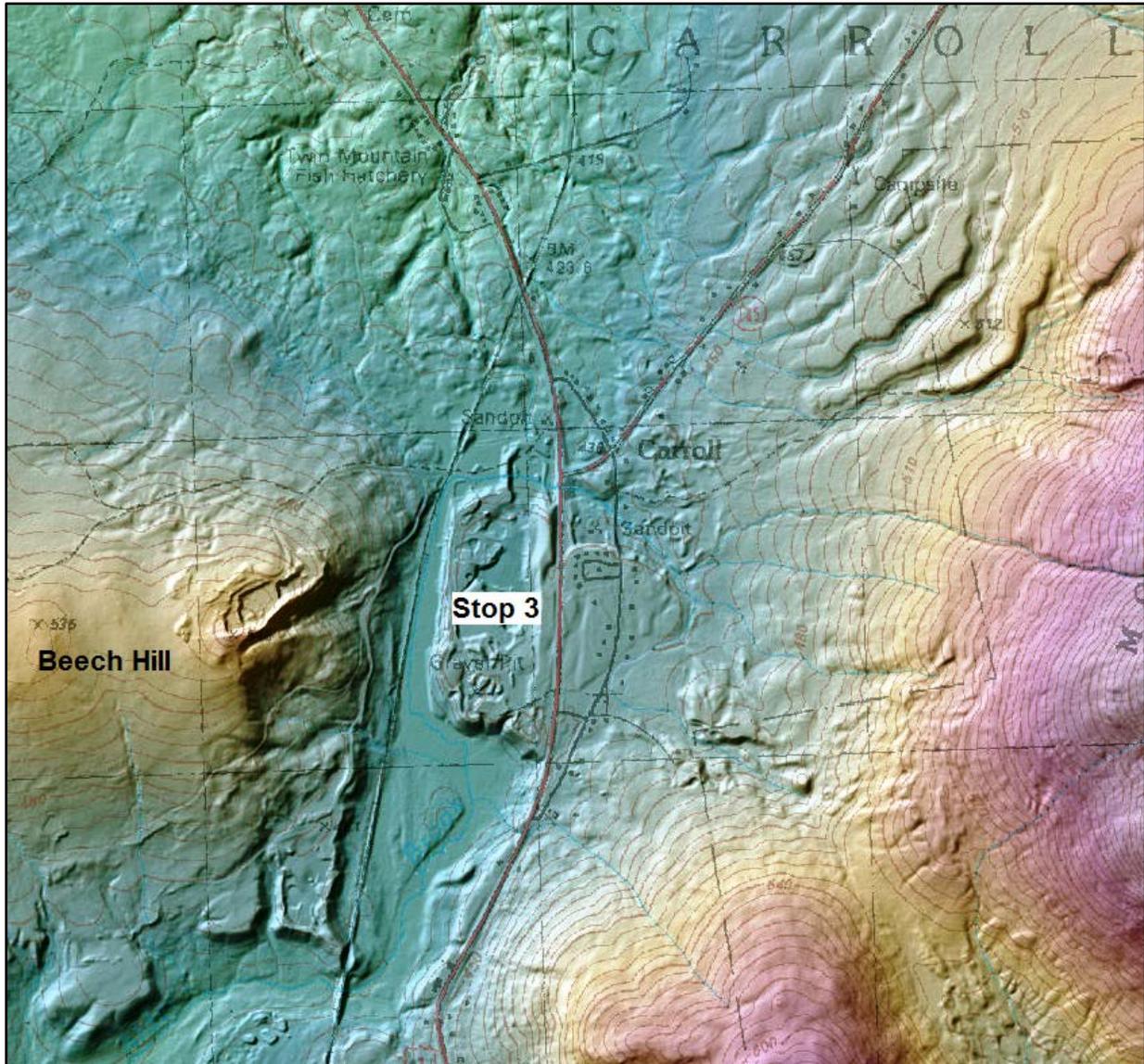


Figure 9. Hillshade LiDAR image (“sun” angle from NW) with color DEM overlay, covering part of the Bethlehem quadrangle. Image shows gravel pits in the E and W parts of the Carroll Delta (center and lower-L), meltwater feeder channels (upper-R), and other features discussed in text. Pit on Beech Hill is a rock quarry.

The coarse gravel forming the delta topset beds can be seen along the upper east wall of the pit, and sandy foreset beds have been exposed in places during recent excavations. In 2000, Carol Hildreth observed some remarkable features at the proximal margin of the delta in the northern part of the pit area. She noted thrust faults and overturned folds in the deltaic sediments, along with north-dipping gravel and till layers plastered against the

delta. Hildreth described these features in detail during the 2002 Friends of the Pleistocene trip (Thompson *et al.*, 2002). She attributed them to ice shove during a minor glacial readvance. She also observed younger, undeformed lacustrine sediments banked against the readvance till, which most likely were deposited in the earliest phase of glacial Lake Carroll to the north.

Incision of the original Carroll Delta resulted from a drop in lake level when recession of an ice tongue farther down the Ammonoosuc Valley (toward Bethlehem) opened up lower spillways (Fig. 3). A deep meltwater channel cut into the west side of the delta as the lake fell is seen along the railroad track. The elevation of the channel at this point is approximately 429 m. Chris Dorion obtained a sediment core from the channel floor (Thompson *et al.*, 2009a, 2017). He located a favorable site that accumulated organic material once the channel ceased to carry meltwater and found sands on the channel floor containing a late-glacial flora. The basal radiocarbon age from Dorion's core showed that the vegetation grew on the channel margins by 11,430 ^{14}C yr BP (~13.3 cal ka BP). The channel itself could have formed several hundred years prior to this time. It may have initially carried meltwater directly from the ice margin, and then briefly served as an outlet for glacial Lake Carroll just before the opening of Lake Israel in the Cherry Pond area, where we have ages as old as 11,800 ^{14}C yr BP (~13.6 cal ka BP).

- 24.7 Exit pit area and turn R on Route 3.
- 26.7 Turn R at light on U. S. Route 302.
- 32.1 Turn L on Trudeau Road.
- 32.3 Turn R into Bethlehem Earth Materials pit. Follow leaders through pit to upper level.
- 32.6 Park on pit floor and put on safety gear. *Be careful to keep a safe distance from steep active pit faces, which may collapse unexpectedly!*

STOP 4. LITTLETON-BETHLEHEM READVANCE SITE (Bethlehem Earth Materials Pit, Bethlehem).

Good stratigraphic exposures are uncommon in the White Mountain Moraine System. Most of the moraines are composed of bouldery till, which is seldom mined for its own sake. However, a few long-lasting pits have been worked near the distal margin of the WMMS in the Ammonoosuc and Israel Valleys, where the Older Dryas Littleton-Bethlehem readvance built moraines and other till deposits on top of earlier deltaic and outwash sediments. One such area is the vicinity of Trudeau Road in Bethlehem, including Stop 4 and the active gravel pit and other excavations associated with the North Country Environment Services (NCES) landfill SE of here.

The Bethlehem Earth Materials pit has been operated in two principal levels over the past three years, exposing up to ~30 m of total section. The lower level, in the northern part of the pit area, formerly showed ~9 m of pebble to boulder gravel and sand on bedrock. The rock is coarse, massive to weakly foliated, pinkish granite with xenoliths. It is part of an Ordovician gneiss dome belonging to the Oliverian Plutonic Suite (Lyons *et al.*, 1997). This rock is now being quarried to produce crushed stone.

Most of the current pit operation is in the higher level to the south, where up to 20 m of glacial sediment are exposed in the long E-W face. The appearance of the pit face changes as excavations continue and the position of the working face migrates southward, but it generally shows till overlying poorly-sorted pebble to boulder gravel with lesser sand and pebbly sand (Fig. 10). The gravel reaches the surface in the west end of the pit, where the face curves around to the NW. In July 2017, a fresh cut in the E end of the pit showed an excellent exposure of compact, fissile, silty-sandy till overlying coarse, poorly sorted gravel. Well-rounded stones that are present in the till throughout the pit face most likely were incorporated from the gravel during the ice readvance. Shear structures including deformed sand lenses are also common in the till.

Large masses of glacially disturbed bedrock underlie gravel in the western part of the pit, and solid ledge was recently encountered in the pit floor. A considerable amount of bedrock may be exposed by the time of our NEIGC trip. The original ledge surface will be examined for glacial striations when safe to do so.



Figure 10. View looking SE at section in the Bethlehem Earth Materials Pit, showing gray readvance till (in upper face) overlying glacial gravel. The till has abundant shear structures, including deformed sand lenses, and rounded stones incorporated from the gravel. Photo taken in December 2015; from Thompson *et al.* (2017).

This pit and most other pits along Trudeau Road are in ridges that trend ENE-WSW. These ridges are best seen on the old topographic map of the Whitefield 15-minute quadrangle. We interpret them as moraines, though they are not as sharp-crested as many others in the WMMS. The Trudeau Road moraines exhibit a readvance stratigraphy that varies in detail from place to place. Recent exposures at Stop 4 have shown just one principal till unit overlying gravel. On the other extreme, many years of excavations and test borings at what is now the NCES landfill area have revealed a complex and variable interlayering of till with silt-sand-gravel units (see above description of Bethlehem moraine complex for details).

The depositional history of the waterlaid glacial sediments in this area is not well understood. The puzzle is complicated by a complex sequence of events during deglaciation, coupled with the jumble of descriptions from various people investigating numerous sites over a long time. We see glaciofluvial gravels, like those in the Bethlehem Earth Materials Pit, while pits and borings around the NCES site have shown abundant glaciolacustrine silt and sand. A basic question is whether the waterlaid sediments occur as local packages, or do they form a widespread continuous stratigraphy of fluvial and lacustrine deposits? Either way, these deposits were overridden by the Littleton-Bethlehem readvance and in places became part of the moraine ridges.

We propose a series of events as follows. As the ice margin receded from the southern Trudeau Road area, outwash sand and gravel was deposited by meltwater streams flowing southward into the upper Gale River basin (probably to glacial Lake Gale, proposed and named by Hildreth (2002)). The ice soon retreated slightly north into the Ammonoosuc basin, resulting in local ponding of deltaic lake sediments in the NCES landfill area. Gravel and sand in the vicinity of Stop 4 may have been deposited by meltwater flowing into this water body. The lake in turn drained southward into the Gale River valley. Then the ice sheet readvanced across the northern Trudeau Road area, deforming the fluvial/lacustrine sediments and building moraines as it experienced oscillatory retreat into the Ammonoosuc Valley. Meanwhile, the upper part of Lake Ammonoosuc (in the Twin Mountain village area) was cutting successive spillway channels SW between the moraines, including the G5 spillway (Fig. 3) at the NCES site. Water flow along these channels reworked earlier glacial sediments and modified the terrain to some degree.

- 32.9 Exit pit and turn L on Trudeau Road.
- 33.1 Turn L on Route 302.
- 36.1 Turn R (at cemetery) on Prospect St. in Bethlehem village.
- 37.2 Turn L into parking lot at Bretzfelder Park.

STOP 5. GLACIAL LAKE AMMONOOSUC DRAINAGE CHANNELS (Bretzfelder Park, Bethlehem).

If time permits, we will stop at Bretzfelder Park in Bethlehem to consider the multitude of glacial meltwater channels in this area (Fig. 11). A few channels, like the one seen here along Barrett Brook, were direct outlets (spillways) from the Bethlehem stages of glacial Lake Ammonoosuc (Thompson *et al.*, 1999). Many other channels in this area carried the lake outflow farther west and south to the lower Gale and Ammonoosuc Rivers. The latter channels record a complex drainage sequence along the retreating ice margin, while small recessional moraines were also forming here (part of the Bethlehem moraine complex described above).

The general succession of lake spillways and other meltwater drainages in the Bethlehem area has been recognized for a long time, but the number and intricate pattern of these channels (Fig. 11) was apparent only when LiDAR imagery became available in the past year! The succession of channels and their spatial interrelationships support an incremental lowering of the ice-dammed lake and a rapidly changing series of drainage paths as the ice margin retreated. This scenario is consistent with the progressive lowering of glacial Lake Ammonoosuc recorded by the Carroll Delta and younger deltas/terraces around Twin Mountain village. Moreover, it provides further evidence that the ice margin was a tight dam for the lake, rather than “leaking” randomly through englacial drainage.

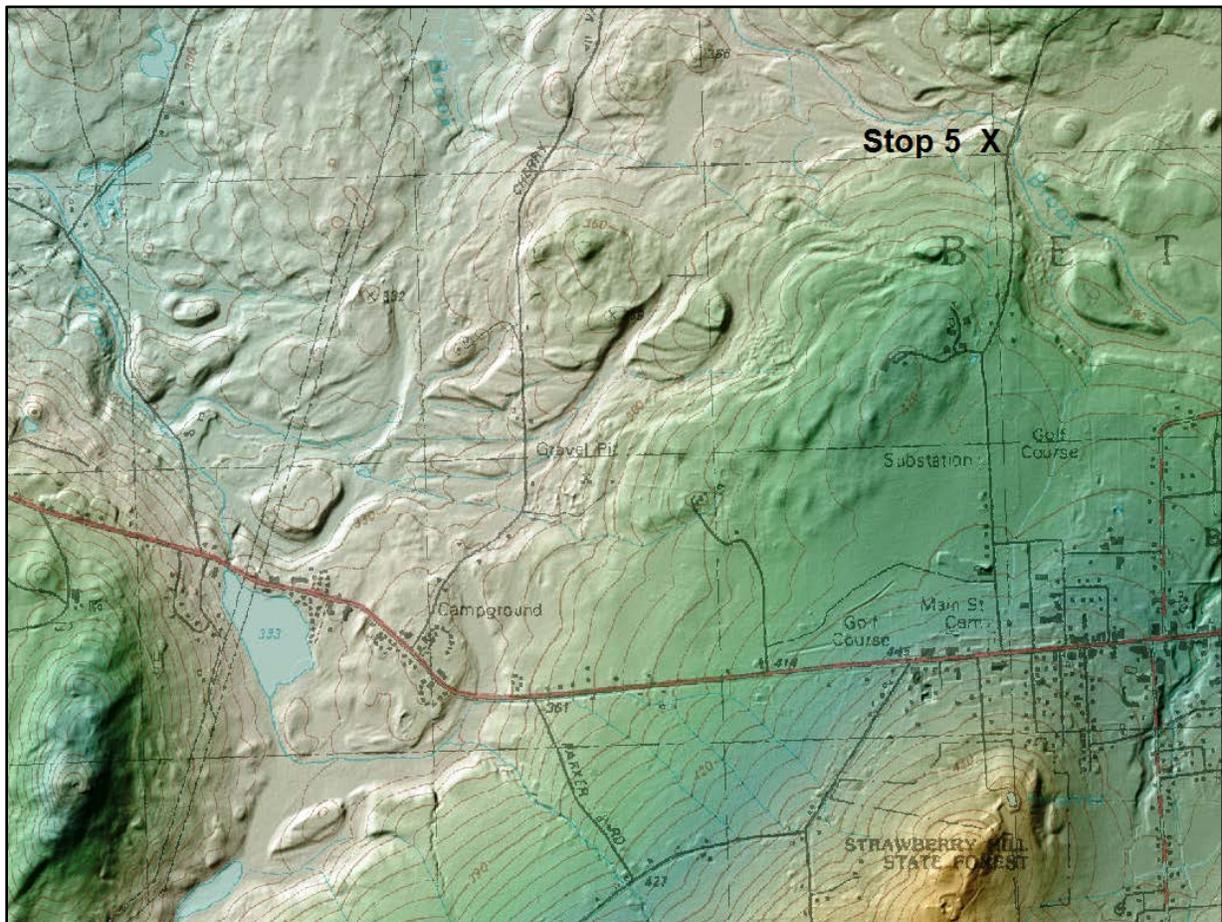


Figure 11. Hillshade LiDAR image (“sun” angle from NW) with color DEM overlay, showing meltwater channels near Bethlehem village.

- 37.3 Exit parking lot and turn R on Prospect St.
- 38.4 Turn L on Route 302.
- 46.8 Turn L at light on Route 3.
- 48.7 Turn R on Route 115.
- 55.2 Turn L on Route 115-A.
- 56.9 Keep R at fork, staying on 115-A.
- 57.6 Turn R onto unmarked private driveway. Continue past house and park in clearing at end of woods road.

STOP 6. MORAINES IN THE ISRAEL RIVER VALLEY, JEFFERSON (Kluckie Pit, Jefferson).

This stop is on the NE side of the Israel River valley, midway between Bowman (Stop 1) and Lancaster. Mapping of the Jefferson quadrangle (Thompson, 2016) delineated a few moraine ridges here, along with curious round to irregular hummocks. In the 1930s, James Goldthwait conducted a gravel survey for the State, and his unpublished map of the area likewise showed the latter features, which he labeled as “till hummocks with blocks [boulders]”. At Stop 6 we can see some of the distinct moraine ridges. These and most other such ridges in Jefferson trend E-W to NE-SW and are associated with the more numerous till hummocks. LiDAR imagery (Fig. 12) shows that the area of hummocky moraine extends farther up the Israel Valley than previously known. The part of the valley floor between the “Old Railroad Grade” and the southeast corner of Figure 12 contains this newly recognized part of the hummocky moraine complex, which had been concealed by forest cover, swampy terrain, and the nondescript contours of modern topographic maps. A SE-trending esker ridge appears in the NW to central part of Figure 12. This discontinuous gravel ridge marks the path of a subglacial meltwater drainage tunnel that probably conveyed sediment to Lake Israel deltas and subaqueous fans in the upper Israel Valley.

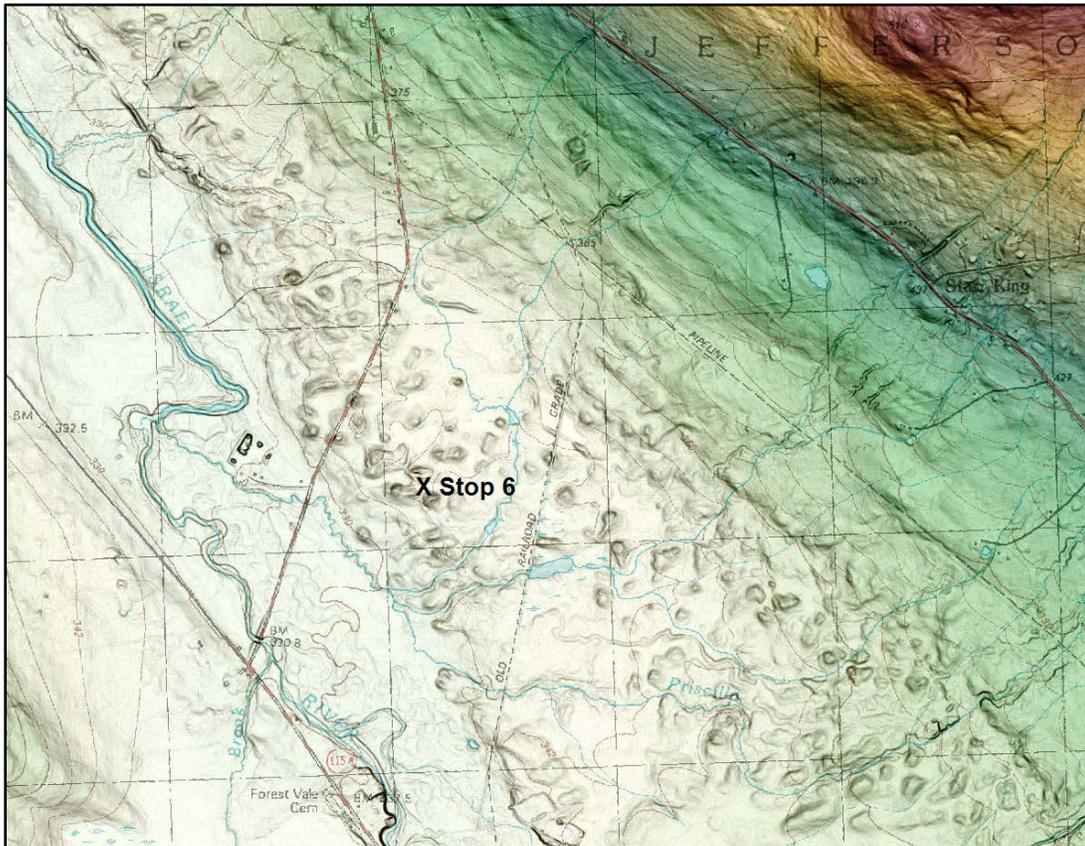


Figure 12. Slope map with color DEM overlay, showing extensive complex of hummocky moraine and short moraine ridges in Jefferson.

Figure 13 shows an excavation in the area where we will park. It recently exposed a nice longitudinal section through a moraine ridge that trends E-W, with a crest elevation of 342+ m. (This section may have slumped considerably by the time of our visit.) The moraine consists entirely of till, with an exposed thickness up to ~5 m. The till is light olive-gray, massive, loose to compact, sandy, and very stony. A few clasts show glacial striations. Various rock types occur here, including many fragments of local bedrock from an Oliverian gneiss dome of Ordovician age (Lyons *et al.*, 1997; Baker *et al.*, 2016). None of the beige or pinkish flow-banded Jefferson rhyolite was seen in this pit, despite its abundance in float boulders and artifacts at Paleoindian sites just NW of here, on the other side of Route 115-A. Other moraine ridges occur in the clearing S of the parking area. We will examine one of these if time permit.

The stony till at this locality resembles that which forms the moraines of the WMMS. It likewise may have been deposited at the margin of the receding ice sheet, but there is little evidence of depositional processes in the presently exposed section. The depositional environment of the nearby hummocky deposits is even more uncertain due to the lack of exposures. Their location is slightly down the Israel valley from the inferred proximal limit of the WMMS, and the hummocky topography is different from that of the moraine belt. We infer that the hummocks record local stagnation of the Israel Valley ice lobe following the Older Dryas readvance that built the Bowman Moraines. The decaying ice probably developed depressions into which flow till slumped from the debris-rich glacier surface. Some of the hummocks lie below the inferred contemporary level (~339 m) of the Baileys Stage of glacial Lake Israel. We expect that excavations of the mounds would reveal mixtures of diamict with fluvial and lacustrine sediments showing collapse structures, of the sort described by Benn and Evans (2010).

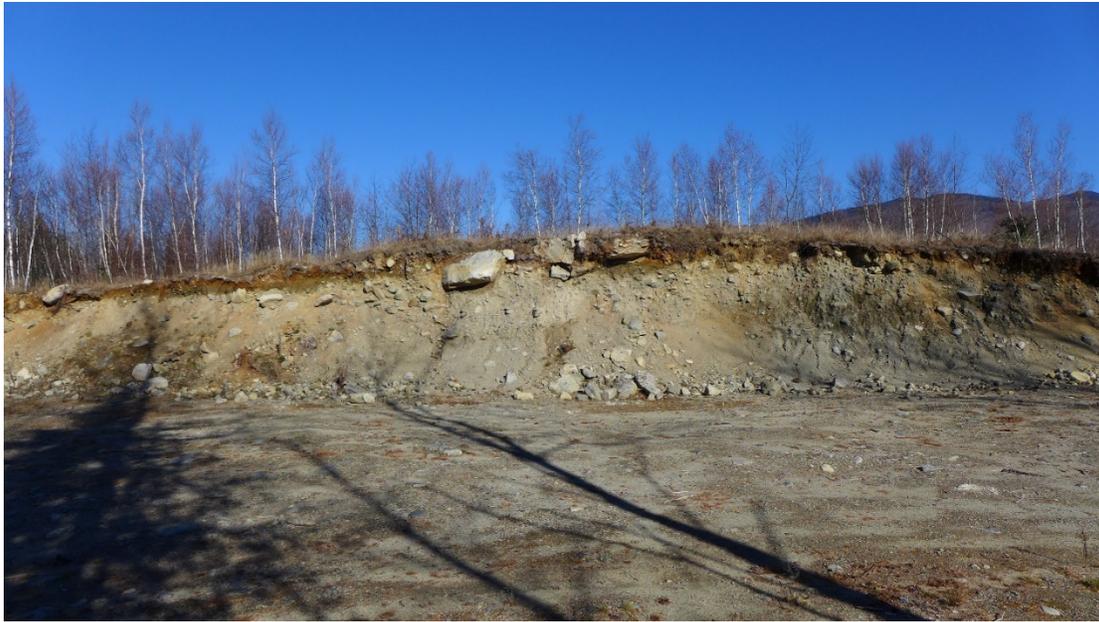


Figure 13. Pit face showing longitudinal section along distal (S) side of a moraine in Jefferson. Photo looks north and was taken in 2015.

- 58.2** Return to Route 115-A and turn R.
- 59.8** Turn L on Route 2 in Jefferson village.
- 60.5** Turn R on North Road.
- 62.7** Turn R on Gore Road.
- 64.0** Sharp bend to L, on what is now Garland Road.
- 64.5** Turn R on Pleasant Valley Road. This road climbs up the gently sloping lower part of the Garland Brook alluvial fan.
- 65.2** Jct. with Arthur White Road. Keep L on Pleasant Valley Road.

65.5 Turn R on Community Camp Road. Follow leaders beyond end of town road and onto the Crane family's private road.

66.9 Park in upper part of large shallow gravel pit.

STOP 7. BUNNELL BROOK ALLUVIAL FAN (P&R Excavating Pit, Lancaster) & SAPROLITE EXPOSURE.

At this stop we will first examine one of the many alluvial fans that have developed on the lower slopes of the White Mountains. The Bunnell Brook fan is a composite feature consisting of an earlier, higher, and poorly understood fan deposit (Qfb₁ on the map by Thompson, 2016), which was incised and followed by the lower fan adjacent to modern Bunnell Brook (Qfb₂). The Qfb₂ fan merges downstream with a large, more gently sloping fan along neighboring Garland Brook to the south. We have not studied these fans in detail, so our observations are preliminary and will hopefully encourage further studies of fan deposits in the White Mountains.

Stop 7 is a broad shallow gravel pit in the lower Bunnell Brook fan (Fig. 14). The pit exposes about 3 m of very coarse, poorly sorted pebble-to-boulder gravel. Many of the gravel stones are somewhat angular, suggesting they have not been transported far by water. Boulders ≥ 1 m in diameter are common here, and a few are up to 3 m and larger. The LiDAR image shows a braided stream channel pattern on the fan surface, supporting a fluvial origin for this deposit.

Farther up the valley, Figure 14 reveals a cluster of arcuate head scarps where landslides have occurred next to Bunnell Brook. A reconnaissance of the slide zone did not reveal fresh sections, but a small recent slide was found southeast of the brook an elevation of ~ 570 m, just north of the former (now closed) Mt. Cabot Trail. The fresh section exposed by this slide showed sandy, stony diamict (probably colluvium) and poorly sorted gravel sharply overlying stratified, water-saturated silt-sand interpreted as a lacustrine unit. This stratigraphy suggests that the large NW-draining basin of Garland Brook and its tributaries may have been occupied by an ephemeral ice-dammed glacial lake during deglaciation of the area. Loading of the lake sediments by glacial or postglacial deposits may be responsible for triggering the Bunnell Brook landslides.

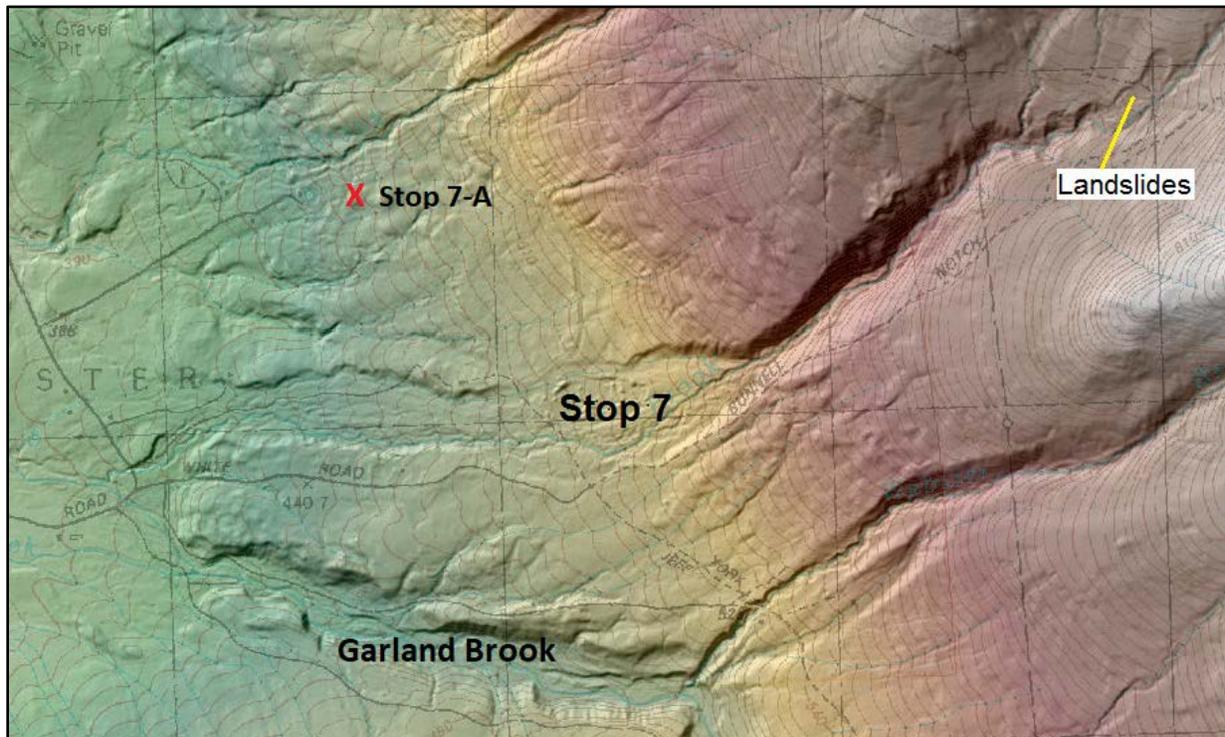


Figure 14. Hillshade LiDAR image (“sun” angle from NW) with color DEM overlay, showing the Bunnell Brook fan and headscarps of landslides along upper part of the brook.

The coarse gravel at Stop 7 resembles many other fan sediments seen in stream banks and pit exposures around the White Mountains. We found very little published research on these fans, unlike the debris avalanches that have received considerable attention in northern New Hampshire. The stratigraphy, mode(s) of deposition, and depositional chronology of the fans are still poorly understood. Gravel exposures, observations of modern brooks, and LiDAR channel patterns on fan surfaces suggest that transport by steep mountain streams was the dominant process, but debris flows may have played a significant role in fan construction. The criteria used by Jackson et al. (1987, p. 116) to identify debris flows in the Canadian Rocky Mountains may prove helpful in the White Mountains:

“Fans were classified as having a debris flow component on the basis of the following field geomorphic and/or sedimentologic criteria: (1) exposures of debris flow diamictos characterized by weak stratification, poor sorting and matrix-supported angular clasts... (2) the presence of debris flow levees or debris flow lobes on fan surfaces...and (3) the presence of oversize (>1m) lone boulders on the fan surface...”

Diamictos have not been observed in the lower Bunnell Brook fan (Qfb₂), although pit owner Paul Crane said that he encountered a buried clay layer in one part of the pit at Stop 7. Landslide scars in the upper Bunnell Brook valley (described above) indicate that slide events likely supplied a lot of sediment to the fan on one or more occasions and over relatively short times, possibly including mud flows and/or coarse debris flows. Shallow roadside exposures in the older fan unit (Qfb₁) did reveal sediments of this type. Along White Rd., Thompson found a cut showing ~1 m of pebbly sand overlying silty-pebbly diamict, and other ditch exposures showed weakly stratified pebbly diamict, silt, and sand. On the northern part of the Qfb₁ fan surface, a ditch near the lower end of Community Camp Rd. (aka Mt. Cabot Rd.) exposes stratified clay-silt, locally overlain by sandy pebble gravel in one place and containing a sand lens in another. The best exposure of debris flow facies that we have seen in the region was at the Drouin Pit, located in a composite landslide/fan that resulted from a huge prehistoric slope failure on the hillside just south of the village. The guide for Trip C3 in the 2009 NEIGC guidebook describes and illustrates the stratigraphy at this locality (Thompson *et al.*, 2009b).

The greatest uncertainty surrounds the depositional history of this and other fans in the White Mountains. Most of them are forested, commonly with roads and houses, and show few signs of major aggradation in historical time. It is tempting to propose that these extensive coarse-gravelly fans formed in large part just after deglaciation, when glacially derived sediments on the mountainsides remained unstable and were not yet anchored by tree cover. This theory is consistent with a study in British Columbia whose authors concluded that “rapid sedimentation during the paraglacial period contrasts sharply with present-day conditions” (Church and Ryder, 1972, p. 3059). However, Jennings et al. (2003) carried out a detailed study of fans in Vermont (a geographically more comparable area!) and documented a long history of episodic deposition spanning the Holocene. The fans described by the latter authors are much smaller, finer-grained, and better stratified than those in the northern White Mountains. Thus they preserve datable soil horizons that yielded a chronology of sedimentation events alternating with periods of stability. Calibrated radiocarbon ages from wood samples near the base of the Eden Mills fan in the northern Green Mountains indicate rapid sedimentation immediately after deglaciation, but also the presence of woody vegetation less than 1000 yr following ice retreat (Jennings *et al.*, 2003, p. 195).

In contrast, Wells and Harvey (1987) described a series of alluvial fans in northwest England that were deposited in response to a single very intense rain storm. The duration of the storm was only about 2.5 hours, yet it generated 13 alluvial fans ranging from debris flows to fluvial deposits. Estimated volumes of these fans range from 50 to 2,380 m³ – smaller than those in Vermont but nevertheless impressive for one brief storm. The affected region is mountainous but lacks tree cover, so the degree of sediment mobilization may have been greater than would occur from a similar storm in the White Mountains.

Stop 7-A. After leaving the gravel pit, we will stop at an interesting road cut on the access road. The cut shows a massive to weakly layered, varicolored, pasty sediment that appears to be a mixture of clay, silt, and sand-size particles. The clay may be very local here, because stony diamict is exposed along the roadside a short distance to the southeast. A sample of this clay-rich sediment was analyzed by Alexander Falster, Research Technologist at the Maine Mineral and Gem Museum in Bethel, Maine. Its chemical composition determined via energy dispersive spectral analysis reveals a high-Al silicate bulk composition with minor Fe, Na and Mg. The high Al content suggests the presence of kaolinite, associated with members of the smectite group (montmorillonite). Fe appears to

be mainly a staining agent such as late goethite and/or hematite. X-ray diffraction analysis is needed to confirm these minerals (A. Falster, pers. comm., July 2017).



Figure 15. Clay exposure at Stop 7-A. The stones on upper part of section probably were introduced by road work.

The clay minerals suggested by Falster’s analysis commonly occur in saprolites formed by weathering of bedrock over long time intervals (Carroll, 1970). Saprolite is uncommon in the glacially-scoured terrain of New Hampshire, though remnants have occasionally been found that survived glacial erosion (Fig. 16). Clayey sediment like we see in this road cut was also encountered in digging a house foundation ~290 m downslope (SW) from here, so it covers a sizable area in that direction. While most of the clay at this locality has an orangey to brick-red color, some of it is black, white, or even pale greenish. These same contrasting colors were noted when digging the foundation hole down the street.

Assuming this is a saprolite, its mode of emplacement is uncertain. The possibilities are: development by in-situ rock weathering, glacial transport, or mass wasting from higher on the mountain. Considering the lack of remnant parent rock or evidence of glacial deposition at this site, we propose the clay was deposited as a mudflow from the mountainside to the northeast. LiDAR imagery reveals now-dry alluvial channels, possible slump scars including a long N-S scarp just north of here, and many curious lobate features suggesting mass-wasting processes. Mobilization of saprolite from the deep Bone Brook ravine seems like a good working hypothesis. There is a scarp at the head of the brook that may indicate landslide activity. Also, the upper part of the Bone Brook drainage would have been more protected from glacial erosion than the area of Stop 7-A, and thus would have been a favorable place to preserve remnants of preglacial saprolite. The bedrock in that area is Jurassic granite (Lyons *et al.*, 1997; D. Eusden, pers. comm., 2017), which is consistent with Falster’s preliminary analysis of the clay.

We have not thoroughly researched saprolite landslides, but some insights were found in a paper by Durgin (1977) on the relationship between landslides and the weathering of granitic rocks. This author notes that granite saprolites are prone to slumps (rotational slides), and he observes that “Ground water drains through the failure surface and relict joints, precipitating iron and manganese” (Durgin, 1977, p. 130). The black streaks seen in the clay at this site are likely Mn oxides. And finally, are there any unweathered stones within the clay unit that were deposited contemporaneously with the clay? If so, they might indicate the source and mode of transport of the saprolitic sediment. Maybe we will find some on today’s trip!



Figure 16. Section in excavated area above rock quarry on the east end of Beech Hill, Carroll (seen in Fig. 9). This remarkable exposure showed in-situ saprolite (freshly scraped area in lower-central part of photo) developed in metamorphic rock. At the top of the orange saprolite, the relict vertical rock foliation was dragged to the left by glacial ice when the thin cover of lodgement till was deposited. Note oriented (and striated) rock clast in the gray till just above shovel blade. Photo by W. Thompson, 14 October 2005. View looking SSW.

END OF TRIP. Return to Bethel, Maine via U. S. Route 2.

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