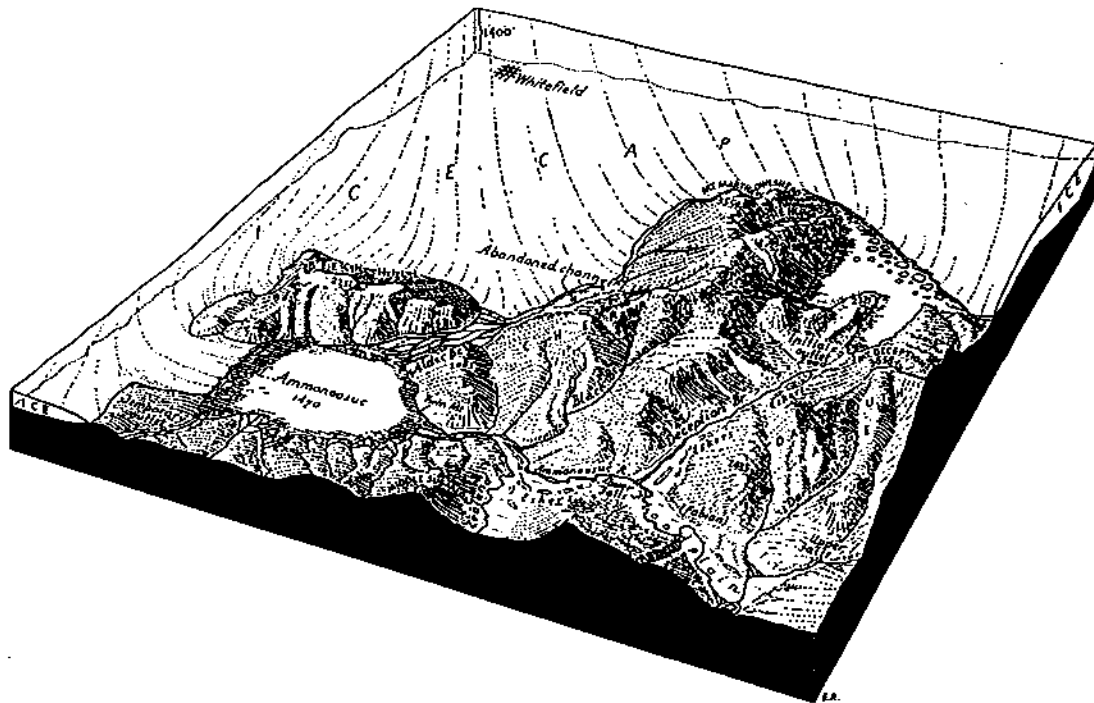


Glacial Geology and Archaeology of the Northern White Mountains, New Hampshire

Guidebook for the
65th Annual Reunion of the Northeastern
Friends of the Pleistocene

Littleton, New Hampshire
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Cover illustration: Diagram from R. J. Lougee (1940), showing location of the late Wisconsinan ice margin when the Carroll Delta was deposited into the Gale River 2 Stage of glacial Lake Ammonoosuc (Thompson *et al.*, 1999). The present location of Twin Mountain village is shown on the eastern side of the lake, and the "temporary outlet" shown near the front-left corner of the diagram is prominently visible today along U. S. Route 3. The Carroll Delta is described under Stop 1 in this guidebook.

GLACIAL GEOLOGY AND ARCHAEOLOGY OF THE NORTHERN WHITE MOUNTAINS, NEW HAMPSHIRE

by

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INTRODUCTION

Welcome to the 65th annual reunion of the Northeastern Friends of the Pleistocene. The trip leaders have developed a personal interest in this part of the Granite State that we would like to share with you. For most of us, this interest was kindled by hiking and camping trips with family and friends. We walked the trails, read the history books, and became fascinated with the geology and archaeology of the mountains. Field work and new discoveries made us realize that much remains to be learned about this region. Some of our findings were included in the guidebook for the 1996 New England Intercollegiate Geological Conference, and most recently in the special White Mountain issue of the CANQUA journal, *Géographie physique et Quaternaire* (Thompson *et al.*, 1996, 1999). Much of the present guidebook is derived from those publications, to which we have added new information.

Figure 1 shows towns and other geographic landmarks mentioned here. The Mount Washington 1:100,000-scale topographic map provides excellent coverage of the entire White Mountain region.

The White Mountains have a long history of geologic research. For over 150 years, scientists have been attracted to this scenic area to study its glacial deposits. Their pioneering work in the mid 1800's occurred during a time of settlement and economic development in the mountain region, including the growth of a major tourism industry. Fundamental concepts and problems in New England glacial geology were tested and debated here.

Two controversies dominated glacial studies in the White Mountains from the late 1800's until 1940. First was the scale of glaciation that occurred in the mountains. Were they affected mainly by alpine glaciers, a regional ice cap, an even larger continental ice sheet, or some combination of these types of glaciers? Once it was realized that continental glaciation was the most important process, there was a heated argument regarding the mode of deglaciation. Was there a distinct northward-retreating ice margin, presumably associated with active glacial flow, or did the ice sheet experience widespread, essentially simultaneous stagnation due to thinning over the mountainous terrain? In the mid 1930's the stagnation viewpoint became the dominant paradigm for the next half century.

The question of active vs. stagnant ice retreat continued to be studied elsewhere in New England during the late 1900's. Meanwhile the White Mountains were largely ignored until recent years. This relative lack of interest probably stemmed from a perception that the area's potential for new field research had been exhausted by previous workers. Geologic literature fostered this notion by giving the impression that late Wisconsinan ice stagnated over the White Mountains without leaving end moraines or other clear evidence of its recessional history (e.g. Goldthwait, 1938; Stewart, 1961).

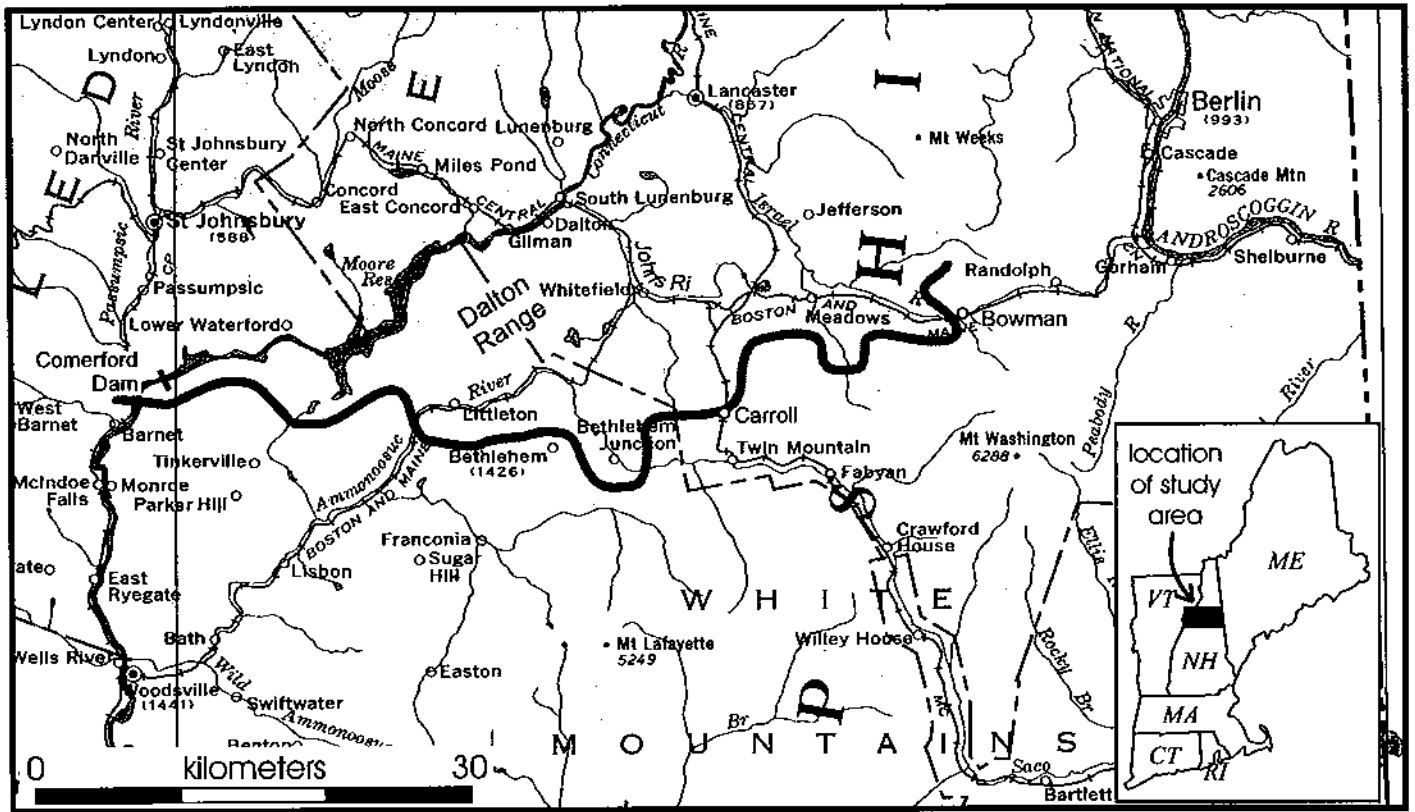


Figure 1. Location map for the northern White Mountains. Thick gray line shows southern limit of the Littleton-Bethlehem Readvance and associated moraines (from Thompson *et al.*, 1999).

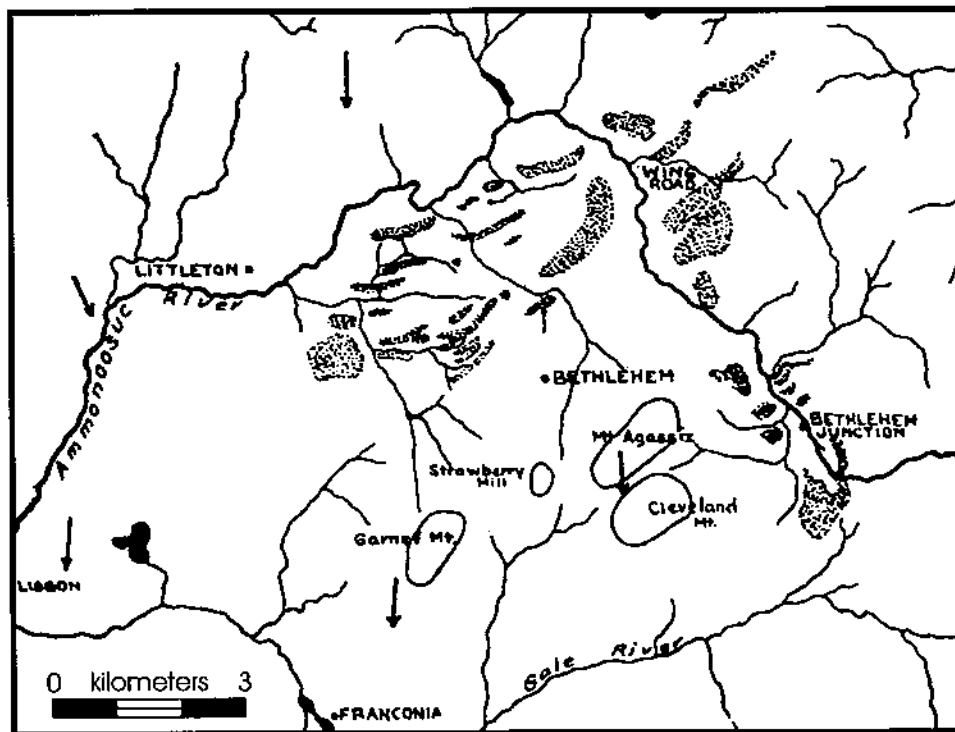


Figure 2. Part of map by J. W. Goldthwait (1916), showing moraines (stippled areas) in Bethlehem, NH. Arrows indicate ice-flow direction at striation localities.

Many of the debates concerning the above problems were focused on the upper Ammonoosuc River basin, extending west from Mt. Washington to Littleton, and the Israel River valley to the north. The present authors have documented the glacial deposits left by ice recession in this part of the White Mountains, including the Bethlehem Moraine complex in the Ammonoosuc basin and the continuation of this complex to the northeast. Our field work indicates the presence of many glaciolacustrine morphosequences like those used to record ice-margin recession elsewhere in New England (Koteff and Pessl, 1981). We also provide evidence supporting the original concept of the Bethlehem Moraine as a complex of end-moraine ridges deposited by active glacial ice. New radiocarbon dates are presented here in the context of the regional deglaciation sequence.

The archaeological record north of the White Mountain notches in New Hampshire is sparse, with rather little known due to limited field research. Interestingly, we find the greatest extent of knowledge concerns the earliest time period, the Paleoindian. Nearly two thirds (seven out of eleven) of the identified sites from this era in New Hampshire are in the North Country, and the town of Jefferson contains the greatest concentration. There, on the glacial till slopes on the northern flank of the Israel River Valley, we find a cluster of at least five hunter-gatherer encampments that date to approximately 12,000 years ago. The sites probably represent seasonal occupations by caribou hunters who stalked herds that migrated through the valley along the bed of the former glacial Lake Israel. The tools and debris left at these sites indicate that more than hunting was carried out here, as evidence for bone and stone tool manufacture and clothing manufacture has been documented. The presence of good quality rhyolite (used for the manufacture of points, knives, scrapers, etc.) among the till stones at the Jefferson sites may have been the reason why these sites occur in such a dense cluster. Significantly, the duration of occupation seems to begin and end with the Younger Dryas climatic episode. There is no evidence for any earlier habitations and, once the Holocene began in earnest, these sites were permanently abandoned. The Paleoindian sites in Jefferson represent a rare, and possibly unique, concentration of late Pleistocene human habitations in the Northeast.

ACKNOWLEDGMENTS

The authors thank the Billings Fund of the Mount Washington Observatory for providing a grant that enabled radiocarbon dating of sediment cores, and for a generous discount on copies of *Géographie physique et Quaternaire* from the Observatory's bookstore. Surficial geologic quadrangle mapping in the Littleton-Bethlehem area was funded by the USGS STATEMAP program in cooperation with the New Hampshire Geological Survey (David Wunsch, State Geologist). The Maine Geological Survey (Bob Marvinney, State Geologist) provided van transportation for this trip; and Brian Fowler (North American Reserve, Laconia, NH) supplied copies of our recent research articles from the Mount Washington Observatory bulletin (*Windswept*). Dave Govatski (White Mountain National Forest, Bethlehem office) and Dick Rush (Gilead Farm, Dalton) helped with contacting land owners and other trip logistics. John B. Poisson, of the Maine Geological Survey's Resource Data Services Division, assisted with preparation of digital graphics for this guidebook. Helpful suggestions for conference logistics and mailing lists from previous Friends trips were provided by Ernie Muller, Julie Brigham-Grette, and Serge Occhietti.

The research conducted at the Jefferson archaeology sites since 1995 would not have been possible without the phenomenal efforts of the SCRAP program volunteers. The Archaeological Conservancy played a pivotal role in the protection of the Jefferson II Site through their timely purchase of the property in 1999. This effort was made possible by the cooperation of the Nevers family, the previous owners, who not only postponed sale of the site to the public once they learned of its importance, but also reduced the price to well below market value to facilitate the sale to the Conservancy. Finally, we express our sincere appreciation to all the residents of Jefferson who allowed a

horde of enthusiastic archaeologists to come into their community and onto their land and carry out this exciting research.

We are very grateful to the individuals and companies listed below for permission to visit their properties, both for mapping and research purposes, and to include them in the 2002 Friends of the Pleistocene trip. Without their cooperation and the assistance of many other landowners, our field projects would have been much more difficult.

Twin Mountain Sand & Gravel (Roger Martin, Pike Industries), Carroll (Stop 1);
 Ken W. Corrigan, Inc., Randolph (Stop 4);
 Archaeological Conservancy (Stop 5);
 Patricia Bacon (Stop 6);
 Weeks State Park Association (Stop 7);
 Rodney Marvin (Stop 8);
 Doug Ingerson, Jr. (Chick's Sand and Gravel; Stop 9);
 Peter Lavoie (Stop 10);
 The Society for Protection of New Hampshire Forests, Bethlehem (lunch stop at Bretzfelder Park);
 Bethlehem Water District (Stop 12)
 John Wedick, Jr. (Wing Road Commons, Ltd; Stop 13).

Many other residents of the White Mountain region kindly assisted us in various ways.

Note: Nearly all of the stops in this guidebook are located on private property. Some of them are working gravel pits, and visitors may not be accommodated due to safety concerns. Ownership and operational status of sites are subject to change. Permission was granted to visit them during the 2002 Friends of the Pleistocene trip, and must be obtained by other persons or groups for subsequent visits!

CONTINENTAL GLACIATION IN THE WHITE MOUNTAINS

The following summary is from Thompson (1999), who reviewed the history of glacial studies in the White Mountains. Systematic investigations of glacial features in this region began in the 1840's, when Charles T. Jackson conducted the first State-sponsored geological survey of New Hampshire. The concept of glaciation was still new at that time, and was not quickly accepted here. In fact, Jackson (1844, p. 24) proclaimed that "the glacial theory of drift is absurd!" The sediments that we now recognize as glacial were attributed by Jackson to a marine flood carrying icebergs. Similar views were held by other prominent scientists, including Edward Hitchcock (1841) and Sir Charles Lyell (1868). Soon after his arrival in the United States in 1846, Louis Agassiz traveled to the mountains with a group of Harvard University students to look for evidence of glaciation (Lurie, 1988). He found moraines and other features which indicated that the region had been covered by glacial ice (Agassiz, 1870).

The glacial theory was more widely accepted in New Hampshire by the 1860's. Alpheus Packard, Jr. (1867a,b) and George Vose (1868) both proposed that a central ice cap in the White Mountains fed a radiating series of valley glaciers. Vose recognized that there had been a more widespread southward glacial flow across New England, but he did not know its timing relative to the supposed local glaciation. He also noted the NE-SW trending striations in the Peabody River valley between Pinkham Notch and

Gorham, but interpreted them as supporting the downvalley flow of a glacier originating in the vicinity of the Presidential Range. Charles Hitchcock (1878) disproved this theory by finding examples of stoss-and-lee bedrock topography showing that continental ice flowed southwestward *up* the Peabody Valley. Some of the outcrops examined by these workers can still be seen along Route 16 near the entrance to the Mt. Washington Auto Road.

Hitchcock became the second State Geologist of New Hampshire in 1868, and carried out an extensive survey of the state over the next ten years. Among his many contributions, Hitchcock found evidence that the summit of Mt. Washington had been glaciated (Hitchcock, 1876). Both he and Agassiz (1870) thought that an ice cap persisted in the White Mountains after the recession of the last continental ice sheet. They claimed that a local glacier flowed northward from the Franconia Range and deposited the moraines at Bethlehem in the Ammonoosuc River valley. Warren Upham, who had worked on the Hitchcock survey, carried the ice-cap model even further when he proposed that a belt of moraines could be traced around the full circumference of the White Mountains (Upham, 1904). James W. Goldthwait (1916) refuted the ice-cap theory by marshaling field evidence, including the provenance of erratic boulders and the distribution of moraines and glacial-lake deposits, to show that continental ice flowing from the north was the dominant type of glaciation.

Most early investigators, such as Packard (1867a,b) and Hitchcock (1878), assumed that dynamic ice persisted in the White Mountains during deglaciation. They cited many examples of what they thought were end moraines deposited during the recession of local glaciers. After the importance of continental glaciation was established, Goldthwait (1925) still envisioned a northward-receding ice margin that remained active and deposited moraines. However, in the 1930's there was a fierce debate over the relative importance of active vs. stagnant ice during glacial recession from the White Mountains. This debate often seemed to be carried on for its own sake, rather than leading to any broader conclusions about the history of the Laurentide Ice Sheet. Richard Flint's (1930) work in Connecticut was a major influence that swayed many people toward the stagnation camp (R. P. Goldthwait, 1939).

Goldthwait (1938) eventually abandoned the concept of active-ice retreat in New Hampshire and discredited many of the moraines and other features described in his 1925 volume. Based on his recent survey of gravel deposits for the State Highway Department, he now stated that in the hilly interior region "Recessional moraines, if present at all, are mere scraps, not in line. Local groups of sharply defined ridges, formerly thought to mark halts in the retreat of the margin of an ice lobe, seem now more likely to be crevasse fillings, formed at unknown distances within the wasting icefield and wholly without seasonal or climatic significance. ... As for the valley deposits generally, ice contacts appear on any or all sides of them" (Goldthwait, 1938, p. 347).

The controversy over the style of deglaciation in the White Mountains peaked around 1940. In the face of growing opposition, Richard Lougee (1940) continued to defend the active-ice deglaciation model. Ernst Antevs (1939) and Douglas Johnson (1941) tried to show that the "normal retreat" and "downwasting" models were overly simplistic in their portrayal of a receding ice margin in high-relief terrain. Nevertheless, the concept of wholesale ice stagnation in New Hampshire prevailed through the mid 1900's (Goldthwait *et al.*, 1951) and has only recently been questioned once again in the northern part of the state.

RECENT GLACIAL STUDIES

Detailed surficial geologic mapping in New England during the last few decades has demonstrated that the late Wisconsinan ice margin receded systematically, and that live ice likely persisted over much

of the region (Koteff and Pessl, 1981). This growing awareness has led to reexamination of glacial deposits in the White Mountains. Several authors have found end moraines and other features supporting the presence of late-glacial active ice in some parts of northern New Hampshire, especially where valleys were favorably oriented for sustaining the flow of the thinning ice sheet.

Gerath described the Success and Copperville Moraines in the upper Androscoggin River basin as having formed along an ice margin that was still active during the initial phase of deglaciation (Gerath, 1978; Gerath *et al.*, 1985). The Success Moraine includes lacustrine sediments deposited between the ice margin and the northwest side of the Mahoosuc Range, and the lake deposits are locally deformed and overlain by till. The Copperville Moraine, on the other hand, is the collapsed head of a fluvial ice-contact morphosequence that was deposited on a drainage divide at the head of the Dead River valley northwest of Berlin (Gerath, 1978).

The Androscoggin Moraine, located on the state line east of Gorham, was mapped in detail and described by Thompson and Fowler (1989). It was found to be a much larger moraine complex than originally described by George Stone (1880, 1899). The moraine consists of at least 21 sharp-crested bouldery ridges that are up to 30 m high. These individual segments form arcuate clusters extending across the Androscoggin River valley. They are composed of glacial diamict, locally interbedded with silt, sand, and gravel. The Androscoggin Moraine suggests vigorous late-glacial ice activity in the northeastern White Mountains. It probably was deposited when the Connecticut Valley lobe spilled around the north side of the Presidential Range and into the Androscoggin basin (Thompson and Fowler, 1989). However, Goldthwait and Mickelson (1982) interpreted meltwater deposits west of Gorham (in the area of Stop 4 on this trip) as indicating the importance of downwastage and stagnation during deglaciation of this part of the mountains.

Possible end moraines and complex sequences of interbedded tills and meltwater deposits have long been known in the Peabody River valley south of Gorham. Packard (1867a, p. 238) remarked that "In passing from Gorham, N.H., to the Glen House we see on each side of the road, fine examples of true glacial moraines ... These moraines, presenting vertical cliffs from fifty to one hundred feet high, of clay and mud and gravel, are mixed in confusion ...". The most recent work on the Peabody River sections identifies these deposits as a sequence of Illinoian and late Wisconsinan tills and associated glaciolacustrine sediments formed in ice-dammed lakes that occupied the north-sloping Peabody Valley (Fowler, 1999).

Research during the past decade has supported the active-ice deglaciation model favored by Crosby (1934), Lougee (1940), and at first by J. W. Goldthwait (1916, 1925). Thompson *et al.* (1996, 1999) validated many of Lougee's observations in the area of this field trip. Quadrangle mapping in the Littleton-Bethlehem area by Hildreth (2000, 2001) and Nelson and Thompson (in press) has provided a clear picture of the moraine segments comprising the Bethlehem Moraine complex. Other new studies have related the deglaciation of the White Mountains to late-glacial climate and the evolution of glacial Lake Hitchcock in the nearby Connecticut River valley. Ridge *et al.* (1996, 1999) greatly expanded our knowledge of the Lake Hitchcock varve record and its relationship to the deglaciation of surrounding areas. Recent analysis of the vegetation sequence and radiocarbon dating of lake sediments in the White Mountains have been carried out by other investigators (Miller and Thompson, 1979; Davis *et al.*, 1980; Spear, 1989; Miller and Spear, 1999; Cwynar and Spear, 2001). These studies provide important data for reconstructing the climate history of the White Mountains and its impact on the Paleoindians and later settlers. One of the major findings is the record of the Younger Dryas climate oscillation that has been discovered in lake sediment cores (e.g. Thompson *et al.*, 1996; Cwynar and Spear, 2001).

TILL DEPOSITS

Till, generally of the ubiquitous ground moraine variety, is the major unconsolidated surface material in the area. It occurs as compact silty lodgement till and relatively loose sandy ablation till; and it underlies most other surficial deposits. These till deposits typically consist of light olive-gray to dark-gray, nonsorted to poorly sorted mixtures of clay, silt, sand, pebbles, cobbles and boulders. In many places they contain lenses of sand and gravel. However, some of the till was deposited in water and has a crude stratification. This variety is referred to here as waterlaid till.

Recent quadrangle mapping by Hildreth (2000, 2001) has shown that in some parts of the Bethlehem area a fissile, compact till overlies sandy till. Good exposures of this type occur along South Road in the Bethlehem West quadrangle. Where the road crosses the 450-m contour, 2 m of compact till overlies unweathered sandy till. Deposition and compaction of the upper till unit probably occurred during the Littleton-Bethlehem Readvance, based on the fact that the underlying sandy ablation till is not weathered.

In a few areas, good exposures in roadcuts and pits (e.g. on Mount Agassiz and around the Black Brook valley) reveal a boulder-rich till composed of light olive-gray to reddish oxidized, nonstratified, loose, sandy, stony material that has a matrix composed predominantly of grüss (an accumulation of angular, coarse-grained fragments resulting from the disintegration of crystalline rocks). Both the boulders and the grüss materials are whitish or pinkish granite and gneissic granite from the Alderbrook Pluton (Moench *et al.*, 1995; Lyons *et al.*, 1997). The boulder till in the Black Brook valley is well-exposed in pits; it is inferred to be morainal material, similar to that found in the Beech Hill moraines (Stop 2).

GLACIAL LAKE AMMONOOSUC

The Bethlehem Moraine complex and a series of glacial lake deposits are among the most significant features resulting from deglaciation of the northwestern White Mountains. The glacial lakes will be discussed first, since they occupied large parts of the north and west-draining valleys in this area, and much of the Bethlehem Moraine complex was deposited in the Ammonoosuc Valley lake.

We are not certain who was first to recognize the former existence of glacial lakes in this part of the White Mountains. It may have been Warren Upham, whose 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). This was the water body to which Goldthwait (1916) gave the name "Lake Ammonoosuc". It resulted from damming of the 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 identified two levels of Lake Ammonoosuc: a higher level into which the "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 was deposited. He also made the important observation that a glacial lake in the Ammonoosuc Valley would have been dammed by the "Canadian ice-sheet" retreating to the northwest, and not by the margin of a local ice cap withdrawing into the mountains to the south (Goldthwait, 1916).

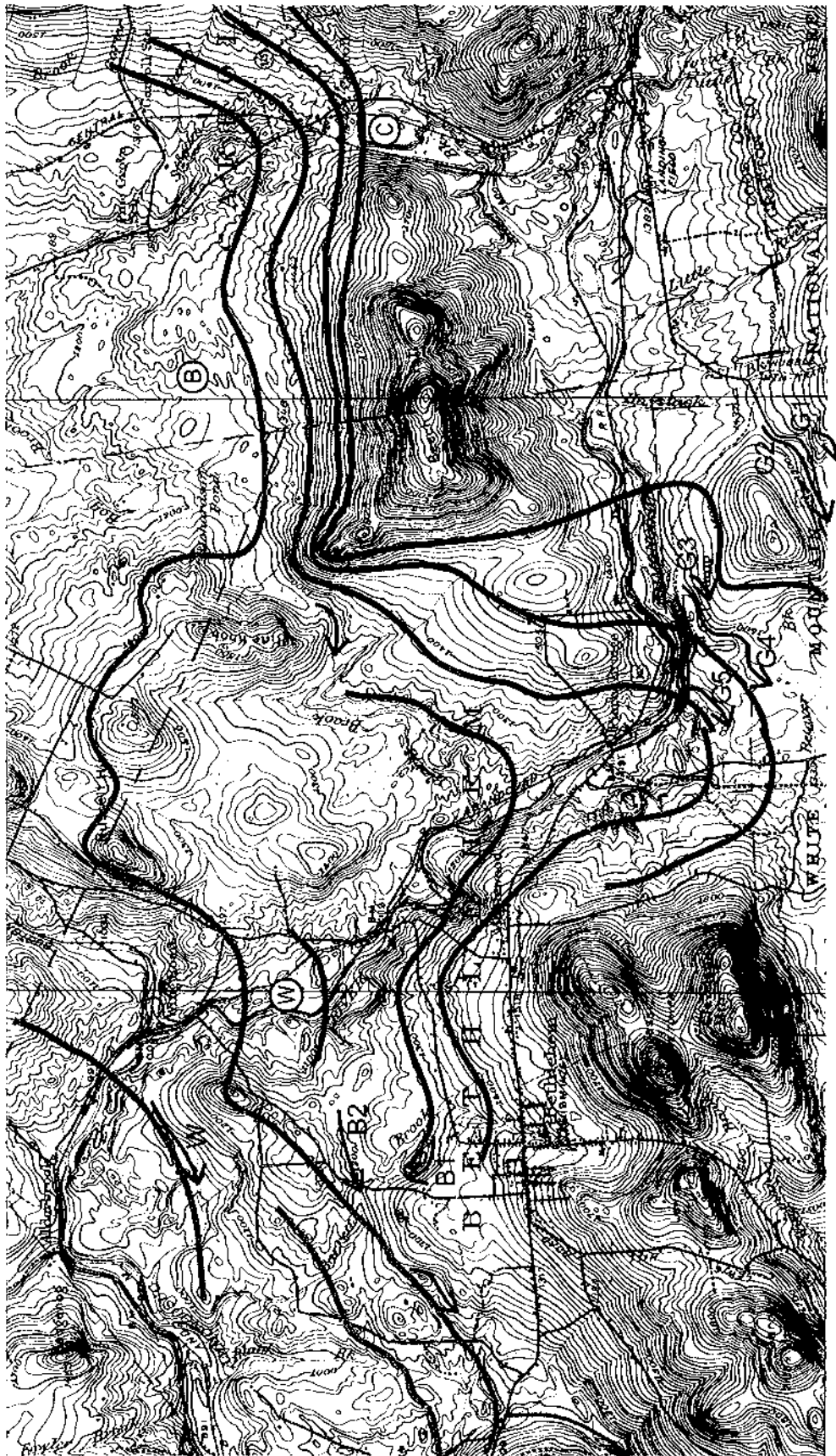
In 1930, Richard Lougee assisted J. W. Goldthwait in the gravel inventory funded by the New Hampshire Highway Department. Lougee was assigned to map several 15-minute quadrangles in the White Mountains. He prepared a manuscript that included a wealth of new information on glacial deposits, lakes, and 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 contained the first analysis of the stages of Lake Ammonoosuc and corresponding spillways. Lougee realized that the earliest stage of the lake (his "Crawford Stage") drained eastward through Crawford Notch. This was the same route that had been taken by a subglacial tunnel drainage which formed the esker in the upper Ammonoosuc Valley (Goldthwait and Mickelson, 1982). The spillway at the notch has an elevation of approximately 573 m (1880 ft).

Thompson *et al.* (1996,1999) located and named the succession of younger stages and spillways for Lake Ammonoosuc that are shown in Figure 3. However, it was Lougee's paper (ca. 1930) that first identified some of these lake levels and their outlets. Following 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 west of Twin Mountain. 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 generally were not very deep. The widest and deepest stage may have been Gale River 2. The log for well CFW 53, located just west of Twin Mountain village, shows a contact between thick glaciolacustrine clay and the underlying till at an elevation of 401 m. Comparison with the nearby G2 spillway elevation of 445 m indicates a water depth of at least 44 m. A geologic cross section of the surficial deposits in this area is shown by Flanagan (1996).

Figure 3 also shows recessional positions of the late Wisconsinan ice margin that correlate in time with the Gale River and later stages of Lake Ammonoosuc. Thompson *et al.* (1996, 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 the known deltas and spillways. The closely constrained relationships between elevations of deltas and terraces in Carroll, and the matching spillways of the Gale River 2-4 Stages, suggest that the receding glacier margin was a tight dam for the lowering lake. Moreover, it appears that the ice sheet had a gentle surface gradient where it wrapped around Beech Hill. Lougee (1940) published the elegant block diagram reproduced on the cover of this guidebook. It shows the ice margin lying against the northwest flanks of Beech Hill and Cherry Mountain when the Carroll Delta was built into Lake Ammonoosuc. This delta is described under Stop 1. (The tiny initials "E.R." in the lower right corner of the diagram suggest that it was drawn by Erwin Raisz of Harvard University, who was known for his skill as a geological illustrator).

During the evolution of Lake Ammonoosuc, water entered 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's diagram. It is likely that much of the sediment deposited in Lake Ammonoosuc came from these non-glacial sources. The mouth of the river shifted farther westward as the lake level dropped, and the lake ultimately disappeared when the ice margin receded from the Alderbrook area in northernmost Bethlehem. At this time, the upper reach of the Ammonoosuc River joined with the lower part of the river that flows southwest from Littleton. Following the drainage of Lake Ammonoosuc, flood plain and alluvial fan deposits have accumulated on the old lake floor. In some places, such as the Little River valley west of Twin Mountain village, these deposits are very thick (Flanagan, 1996).



Partie de la carte topographique de Whitefield, montrant la topographie du complexe morainique de Bethlehem (zone diffuse de crêtes et de buttes dans la vallée de l'Ammonoosuc au nord et à l'est de Bethlehem) ; positions généralisées des marges glaciaires (traits gris) ; chenaux d'écoulement d'eau de fonte (flèches). B : Moraine de Beech Hill ; C : delta de Carroll ; W : balastière de Wedick. Les flèches identifiées (G1, par ex.) montrent les déversoirs aux différents stades du Lac Ammonoosuc. L'altitude des déversoirs est fondée sur la nouvelle carte métrique à 1/25000 de Bethlehem. Stades du lac glaciaire et altitudes : Gale River 1 (G1) : 477 m ; Gale River 2 (G2) : 445 m ; Gale River 3 (G3) : 435 m ; Gale River 4 (G4) : 423 m ; Gale River 5 (G5) : 405 m ; Bethlehem 1 (B1) : 387 m ; Bethlehem 2 (B2) : 369 m ; Wing Road (W) : 327 m.

FIGURE 3. Part of the Whitefield 15-minute quadrangle, showing: topography of the Bethlehem Moraine complex (diffuse area of ridges and hummocks in Ammonoosuc Valley north and east of Bethlehem); generalized ice-margin positions (gray lines); and meltwater spillway channels (arrows). B: Beech Hill moraines. C: Carroll delta. W: Wedick pit. Labeled arrows (G1 etc.) show spillways for stages of glacial Lake Ammonoosuc listed below. Spillway elevations are based on contours from the newer Bethlehem 1:25 000 metric quadrangle. Lake stages and elevations: 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); Bethlehem 1 (B1): 387 m (1269 ft); Bethlehem 2 (B2): 369 m (1210 ft); Wing Road (W): 327 m (1073 ft).

GLACIAL LAKES NORTH OF LAKE AMMONOOSUC (Lakes Carroll, Israel, and Whitefield)

Several glacial lakes - most of them younger than Lake Ammonoosuc - developed as the late Wisconsinan ice margin receded north from Beech Hill and Cherry Mountain. These lakes have not been documented as thoroughly as Lake Ammonoosuc, but we have inferred their history based on the distribution of spillways and sand and gravel deposits that were built into successive lake stages. The reader is referred to Figure 1 and the Mount Washington 1:100,000-scale topographic map for locations mentioned in the following summary of glacial lake evolution in the Whitefield-Jefferson-Lancaster area. Elevations of lake levels and spillways have been interpolated from contours on recent 1:25,000-scale metric quadrangles, and may differ slightly from elevations given in older literature. Flanagan (1996) has compiled data on the distribution, texture, and thickness of the glacial lake sediments.

Ice receding from the north side of Beech Hill dammed a small lake that was called the "Carroll Lake" by Lougee (n.d.), but is here named "Lake Carroll" (in accordance with how other glacial lakes in New England are usually named). Lake Carroll may have very briefly drained south through the 429-m channel that incises the Carroll Delta (Stop 1), but soon would have spilled through the 387-m col just northwest of Beech Hill and into one of the Bethlehem Stages of Lake Ammonoosuc (the higher of which was close to the elevation of the 387-m col) (Figure 3).

The 387-m stage of Lake Carroll is inferred to have been coeval with the highest level (Bowman Stage) of glacial Lake Israel northeast of Cherry Mountain. This stage of Lake Israel drained to the east, across the 457-m divide at Bowman between the Israel and Moose River valleys. Reconnaissance by W. Thompson has identified deltaic and subaqueous fan deposits that were built into the Bowman Stage near Stop 4. As soon as the ice margin withdrew from the north end of Cherry Mountain, Lake Israel could have dropped and perhaps briefly merged with the 387-m level of Lake Carroll. At this time, the glacier margin probably extended from Kimball and Osburn Hills in Whitefield generally eastward to the Jefferson Highland area on U. S. Route 2.

Slight additional ice retreat from Cherry Mountain probably led to the creation of two separate lakes, which were first proposed by Lougee (n.d.). The eastern lake was Lougee's "Baileys Stage" of Lake Israel. Modern topographic maps suggest that the Baileys Stage spilled west across a low 339-m divide at Cherry Pond in Jefferson, and thence into glacial Lake Whitefield (see below). This stage of Lake Israel terminated when escaping lake waters cut channels across the hillside just south of Lancaster. Lake Israel then merged with glacial Lake Coos in the Connecticut River valley. Lougee (n.d.) named Lake Coos, which extended from up the valley from Dalton to North Stratford.

The western lake was called "Lake Whitefield" by Flanagan (1996), and previously named "Johns Lake" by Lougee. It occupied the Johns River valley as the ice margin receded north from Whitefield village. This lake drained south through the 324-m spillway south of Burns Pond and into the Ammonoosuc River. According to Lougee (n.d.), the deposits built into Lake Whitefield include subaqueous (fan?) deposits and a "small deep water esker" near Whitefield. A 285-m channel north of the Dalton Range records the final drainage of Lake Whitefield into Lake Coos in the Connecticut Valley.

THE BETHLEHEM MORAINES COMPLEX

The Bethlehem Moraine was first noted by Louis Agassiz (1870). For many years it was the focus of major controversies involving the modes of glaciation and ice retreat in the White Mountains. We will

first give a brief description of the moraine complex, then tell how it figured in debates concerning the glacial history of the region.

The Bethlehem Moraine complex consists of a varied assortment of till and ice-contact sand and gravel deposits, including dozens of hummocks and ridges, located in the Ammonoosuc River basin in Bethlehem and Littleton. The topography of the moraine complex can be seen in the northwestern part of Figure 3. At first glance, one can understand why Upham (1904) called it a “promiscuous morainic belt”! However, closer examination of both old and new topographic maps and inspection in the field reveal numerous subparallel ridges trending east to northeast. These moraine ridges range in height from 3 m to over 30 m, and up to 600 m or more in length. Most of them occur in a belt extending from the town of Littleton east-northeast to the Wing Road district of Bethlehem. A few less distinct ridges are seen farther up the Ammonoosuc Valley in the Bethlehem Hollow and Pierce Bridge areas. Goldthwait (1916) published a map of the part of the moraine complex lying in Bethlehem (Figure 2).

There are few bedrock exposures within the Bethlehem Moraine complex, but striated outcrops immediately to the north and south indicate ice flow directions of 175-185° (Goldthwait, 1916; W. B. Thompson, unpub. data). Thick surficial deposits are typical of the morainic belt, with a variety of sedimentary facies and stratigraphic associations. Records from the New Hampshire Water Well Board commonly indicate 15-30 m, and locally 50 m or more, of till, sand, and gravel in the Ammonoosuc Valley north and east of Bethlehem village. The general succession in this area consists of till overlain by fine to coarse lacustrine sediments (Lake Ammonoosuc deposits), which in turn are overlain by terrace deposits formed as the early Ammonoosuc River expanded downvalley at the expense of the dwindling glacial lake. The moraine ridges are mostly composed of silty-sandy till, locally containing lenses of sand, gravel, and silt. The tops of the ridges commonly are similar in elevation to, or slightly lower than, the levels of Lake Ammonoosuc into which they were deposited. Large boulders of granite and gneissic granite are very abundant on the surfaces of many of the moraine segments. In the Bethlehem area, most of these boulders were derived from the Alder Brook Pluton (Lyons *et al.*, 1997).

Agassiz gave an enthusiastic description of moraine ridges north of Bethlehem village, along what is now Prospect Street: “The lane starting from Bethlehem Street, following the Cemetery for a short distance, and hence trending northwards, cuts sixteen terminal moraines in a tract of about two miles. Some of these moraines are as distinct as any I know in Switzerland” (Agassiz, 1870, p. 164). These particular moraines have not been apparent to other geologists, which probably has contributed to the skepticism of some later authors regarding the entire moraine complex. C. T. Hildreth (pers. comm.) has speculated that Agassiz may have been writing from memory and misidentified the street where he saw the moraines. Agassiz thought that the morphology and boulder provenance of moraines in the Bethlehem area proved that they were deposited by a local glacier flowing north from the Franconia Range.

Hitchcock (1878) agreed with Agassiz’s theory of local ice depositing the moraines from the south. Upham likewise concurred with this theory, and formally named the Bethlehem Moraine (Upham, 1904, p. 12). However, he believed that the glacial source was a larger ice cap that remained over the high mountains after recession of the continental ice sheet. Upham could not distinguish the sharp moraine ridges described by Agassiz, and instead characterized the moraine as being the “continental type”. His description is close to what we have seen in the field: “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). Upham also said of the Bethlehem Moraine (1904, p. 11) that “... in many places, as at Littleton, its accumulations are more massive than at

Bethlehem". This is true of the high bouldery ridges just southeast of the town of Littleton, along U. S. Route 302 (see Stop 10 and Figure 6). Upham was liberal in extending the moraine; he thought it might be possible to trace it all around the periphery of the White Mountains.

Goldthwait (1916) radically revised the interpretation of the Bethlehem Moraine. Agassiz's model was found to be flawed because it proposed a topographically unrealistic ice-flow path, lacked adequate proof of northward erratic transport from the Franconia Range, and was not supported by striation evidence. Goldthwait also pointed to the lack of recessional moraines near Mt. Lafayette as a problem with Agassiz's and Upham's theories. He said that the geometry, provenance, and associated lake sediments favored deposition of the Bethlehem Moraine from the north by the continental ice sheet. However, our observations do not agree with his contention that "the morainic ground is chiefly composed of sands, gravels, and cobbly boulder beds, with the finer sediment predominating" (Goldthwait, 1916, p. 272). There is a lot of sand and gravel in the area resulting from sedimentation in glacial Lake Ammonoosuc, and this material may have constituted much of what Goldthwait saw in road cuts. The exposures that we have seen, while not very plentiful, indicate that the more clearly defined moraine ridges are made of till and waterlaid till with local interbeds of washed sediments.

Antevs (1922) inferred from his Connecticut Valley varve record that a glacial readvance occurred west of Littleton, in the area where Comerford Dam is now located (see Ridge *et al.*, 1996). He suggested that this readvance might correlate with the Bethlehem Moraine. Both Lougee (n.d., ca. 1930) and Crosby (1934) made the same correlation, which seemed to be supported by the two-till stratigraphy that Crosby found at the dam site. These authors further suggested that the ice margin which deposited the Bethlehem Moraine could be traced eastward to the head of the Carroll Delta (Stop 1). From there, it supposedly wrapped around the north side of Cherry Mountain and the Presidential Range, and was correlated with the till deposits that overlie lake sediments in the Peabody River sections just south of Gorham. Recent mapping by W. B. Thompson in the western part of Littleton has located a few moraines and other ice-contact deposits that may support the westerly continuation of the Bethlehem Moraine, including a prominent moraine segment between Walker Mountain and Mullikin Brook. Most of the Bethlehem Moraine complex is slightly younger than the high-level Carroll Delta at Stop 1 (Figure 3), but is probably considerably younger than any ice-marginal deposits at Gorham.

When J. W. Goldthwait became an advocate of wholesale glacial stagnation in New Hampshire, he dismissed the Bethlehem Moraine as simply being a "zone of massive kettled outwash" (Goldthwait, 1938, p. 349). Lougee (1940) emphatically continued to defend the moraine, and he mentioned the segments south of Route 302 near the Bethlehem-Littleton town line (Stop 10; Figure 6). Evidently the "stagnationists" won the argument, since there was little mention of the Bethlehem Moraine in geologic literature for the next 56 years.

Thompson *et al.* (1996, 1999) concluded that the numerous subparallel till ridges are indeed end moraines. This is especially true of the high, robust ridges such as those near Littleton (Figure 6). Together with the deposits and matching spillways of glacial Lake Ammonoosuc, they indicate the progressive northward retreat of a coherent ice margin during late Wisconsinan deglaciation. The Bethlehem Moraine complex records a large volume of sediment accumulation that is consistent with active ice supplying debris to the Ammonoosuc Basin during retreat through the positions mapped in Figure 3. This retreat was probably interrupted by brief stillstands or forward pulses of the ice margin as the moraines were deposited, but we have not seen any evidence of significant readvance. Shear structures in a superb exposure of a moraine ridge at the Dodge Pit in Littleton (illustrated by Thompson *et al.*, 1999) indicate that the ice was at least locally active.

A remaining problem is the south-southwest orientation of striations around the town of Littleton, reported by Hitchcock (1905). This trend is oblique to the orientation of the ridges that form the Bethlehem Moraine complex. We have not located Hitchcock's Littleton sites, but striations around Bethlehem usually trend very close to due south. As Lougee (n.d.) suggested, the Littleton striations may have formed in the western part of the late ice lobe that occupied the Ammonoosuc basin.

We still need to answer the fundamental questions of when and why the Bethlehem Moraine complex was deposited, and what controlled its location. The following discussion of radiocarbon dates places some constraints on the age of these moraines and their possible correlation with climatic events.

DEGLACIATION CHRONOLOGY AND THE LITTLETON-BETHEHEM READVANCE

Note: The ages mentioned here are in radiocarbon years unless otherwise indicated.

Thompson *et al.* (1996,1999) compiled previously published radiocarbon dates from ponds in the White Mountains and adjacent areas, together with four new AMS dates from the region. Most of these dates are from the lower or basal portions of pond sediment cores, and they limit the time of deglaciation of each site. Detailed information on the sediment cores and new dates acquired by Dorion and Thompson was presented during the 1996 NEIGC conference (Thompson *et al.*, 1996). Pond of Safety, in the mountains just north of the Presidential Range in New Hampshire (Figure 5), was cored because it is favorably located to provide a minimum age for deposition of the Androscoggin Moraine to the east and a maximum age for the Bethlehem Moraine complex to the west. The basal date from Pond of Safety is 12,450 +/- 60 BP (OS-7125). This date is compatible with other limiting dates from northern New Hampshire, as well as the date of 12,250 BP (OS-7119) from Surplus Pond in western Maine. The cores from both of these ponds also show sedimentological evidence of a climatic reversal that probably occurred in Younger Dryas time (Thompson *et al.*, 1996).

Additional radiocarbon dates were recently obtained from cores taken in other ponds in the northern White Mountains (Boisvert *et al.*, 2002). They are presented here in Table 1 and discussed at Stops 1 and 3. Together with previous results, they suggest that the Littleton-Bethlehem Readvance occurred (and the Bethlehem Moraine complex was deposited) ca. 12,000 BP (~ 14,000 cal yr). A short distance west of Littleton, Ridge *et al.* (1996) found stratigraphic evidence supporting the hypothesis of previous authors that a glacial readvance occurred in the vicinity of Comerford Dam on the Connecticut River (Antevs, 1922; Crosby, 1934; Lougee, 1935). The latter authors 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 genetic connection with the deposits historically known as the Bethlehem Moraine.

Based on recent work on the glacial Lake Hitchcock varve sequence in the Connecticut River valley, Ridge *et al.* (1999) determined an age of 11,900-11,800 BP for the Littleton-Bethlehem Readvance. Moreover, Larsen (2001) obtained a radiocarbon age of 11,900 BP for the Middlesex Readvance near Montpelier, Vermont. He correlated this event with the readvance in the Comerford Dam area (and presumably with the Littleton-Bethlehem Readvance in general). These ages from similar latitudes in north-central Vermont and the Connecticut River valley, taken together with our observations in the White Mountains, suggest that the Littleton-Bethlehem Readvance and its correlatives to the east and west were a major glacial event across this part of northern New England.

Thompson (1998) and Thompson *et al.* (1999) suggested that the Littleton-Bethlehem Readvance occurred during the Older Dryas Chronozone. This cold interval began ca. 12,000-12,000 BP and lasted only about 200 years (Donner, 1995; Wohlfarth, 1996). The GISP2 ice core from Greenland likewise indicates a brief cold period around 12,000 BP that is equated with the Older Dryas (Stuiver *et al.*, 1995). The ice sheet must have responded quickly in northern New Hampshire if deposition of the Bethlehem Moraine complex was limited to this brief interval.

The radiocarbon dates in northern New Hampshire and adjacent areas show a generally consistent pattern of being younger to the north, in the direction of glacial retreat. If these dates closely approximate the time of deglaciation, they show that the late Wisconsinan ice margin receded from the White Mountains to southeastern Québec between 13,000 and 11,000 ¹⁴C years ago. This time frame is somewhat younger than the 14,000-13,000 BP deglaciation interval proposed for this region by Davis and Jacobson (1985) and Occhietti (1989). Parent and Occhietti (1988, 1999) used shell ages from near the upper marine limit in southeastern Québec to infer deglaciation and marine transgression in the St. Lawrence Lowland north of New Hampshire at ~ 12,000 BP. This raises the question of whether the Québec marine shell dates are too old or the deglaciation dates inferred from pond cores are too young. A correction for the marine reservoir effect or other factors could make the marine shell dates younger, and the terrestrial dates may lag behind the time of deglaciation. See Thompson *et al.* (1999) for further discussion of the regional correlations between the White Mountains and southeastern Québec.

THE ISRAEL RIVER PALEOINDIAN COMPLEX IN JEFFERSON

In October 1995 a windstorm uprooted trees in the northern New Hampshire town of Jefferson. A local resident, who was an experienced volunteer in the New Hampshire State Conservation and Rescue Archaeology Program (SCRAP), seized the opportunity and systematically inspected the tree throws, recovering debitage and the base of a fluted point. He immediately recognized the significance of the find and notified Richard Boisvert at the New Hampshire Division of Historical Resources. Within days the site (27-CO-28/Jefferson I) was mapped and additional debitage recovered (Bouras and Bock, 1997). Over the next five years, survey and excavations by SCRAP identified four more Paleoindian sites (27-CO-29/Jefferson II Site, 27-CO-30/Jefferson III, 27-CO-45/Jefferson IV, and 27-CO-46/Jefferson V). This cluster of sites is defined as the Israel River Complex. It is a highly unusual, and possibly unique, Paleoindian manifestation in northern New England (Boisvert 1998, 1999a; Boisvert and Puseman, 2002).

Central to the significance of the sites is that, to date, there is no evidence for any occupations by later prehistoric people. On three sites, 129 square meters of test pits and small block excavations have been completed, along with more than 250 shovel test pits (each 50 cm square). With the exception of late 19th and 20th century debris in the plow zones and A Horizons, there are no artifacts recovered that can be identified as post-Paleoindian. This circumstance permits inclusion of the manufacturing byproducts and otherwise non-diagnostic chipped stone tool fragments into the Paleoindian artifact assemblage. Later prehistoric sites are known in the vicinity, but their location and suites of lithic raw materials are distinctively different from the Paleoindian sites.

The Jefferson sites are located in the Israel River valley, a tributary to the Connecticut River 15 km to the northwest. The drainage divide at Bowman separates the head of the Israel River from the uppermost reach of the Moose River, which joins the Androscoggin River approximately 25 km to the

east at Gorham. Together these river valleys form one of the very few corridors of travel in an east-west direction through northern New Hampshire.

The micro-environmental setting of the Israel River Complex is extremely unusual among Paleoindian sites in the Northeast. The sites are situated on ablation till rather than the outwash or eolian sands that underlie virtually all other Paleoindian sites of this region. Additionally, the sites are located partially on hillsides that slope from 10 to 20 degrees. This is highly unusual, as sites from any prehistoric period rarely occur on such slopes.

The sites are large in size and contain discrete habitation loci. Shovel test pit surveys and small block excavations have identified areas of comparatively intense activity. At the Jefferson II Site (ca. 4 ha) at least four loci have been identified and two have been investigated. One excavation block appears to have captured most of a habitation locus that covered approximately 50 square meters. Analysis of the tool distribution suggests that distinct areas of fluted point manufacture, bone/antler tool manufacture, and use of delicate retouched flakes, possibly for clothing manufacture, are present. Basal fragments of fluted points made from exotic cherts (recovered in close proximity to channel flakes) and fluting failures made from locally available rhyolite strongly suggest replacement of damaged projectile points. A cluster of ephemeral pits with a greater than normal amount of charcoal is restricted to one edge of the locus and seems to be associated with cutting and scraping tools. A second locus was partially sampled with a nine square meter block and revealed a similar, though obviously attenuated, pattern. At the Jefferson III Site (ca. 2 ha), a ten meter east-west test trench excavated in an eroding logging road intercepted the peripheries of two loci. A total of 12 and 16 square meters were excavated from the east and west loci, respectively. Again, we recovered a mixture of artifact types replicating the varieties from the Jefferson II site. Three more loci have been identified in the Jefferson III site. Similarly, we have identified two loci at the Jefferson I Site (0.5 ha) which are separated by 40 meters. Jefferson IV consists of a single locus which has been heavily impacted by 19th and 20th century domestic and agricultural development. Even so, it has produced significant data in the form of a pair of fluted points and a Munsungun chert flake which retained cervid (most likely caribou) protein (Puseman, 2001). At the Jefferson V site, a pair of loci were identified through shovel test-pit survey, and little beyond its presence can be verified due to the limited amount of field investigation.

The presence of discrete loci at the sites is similar to other Paleoindian sites in the Northeast where concentrations of artifacts occupying equivalent areas are segregated from each other with the intervening space nearly void of cultural remains. The pattern of tool distribution in the Jefferson sites argues for occupation by small groups, likely at the household level. Estimation of the number of groups resident at any one time, however, cannot be determined with the available data.

The dominant lithic raw material at the sites is flow-banded rhyolite. Initially it was identified as Mt. Jasper rhyolite from Berlin, New Hampshire, 30 km to the east. This material occurs in Paleoindian sites in Maine, New Hampshire, and Massachusetts (Pollock, Hamilton, and Boisvert, 1998). Our survey in Jefferson recovered numerous samples of a raw material with the same visual and textural characteristics. However, detailed comparisons by X-ray diffraction indicate that the Jefferson flow-banded rhyolite is discernible from the Mount Jasper rhyolite (S. G. Pollock, pers. comm.), and the similarities between the two may be explained by a common magmatic source. The Jefferson rhyolite is most likely derived from bedrock sources within 5 km of the site. Cobbles of this material can be easily found in nearby stream beds and till deposits. The specific location of the sites in Jefferson may have been influenced, at least in part, by the distribution of this rhyolite. Lithic workshops are present in the Israel River Complex; and analysis of bifaces, biface fragments, and debitage clearly demonstrates that fluted points were manufactured at the sites. All stages of production, from acquisition of raw material through removal of channel flakes and final trimming, were executed here (Boisvert, 1999b). Extensive

use of a locally available raw material at a Paleoindian site is a significant departure from the pattern observed for most of the Northeast. The only other Paleoindian sites in northern New England where locally available material was used are workshop sites near the Munsungun source in northern Maine (Bonnichsen *et al.*, 1981; Bonnichsen *et al.*, 1991).

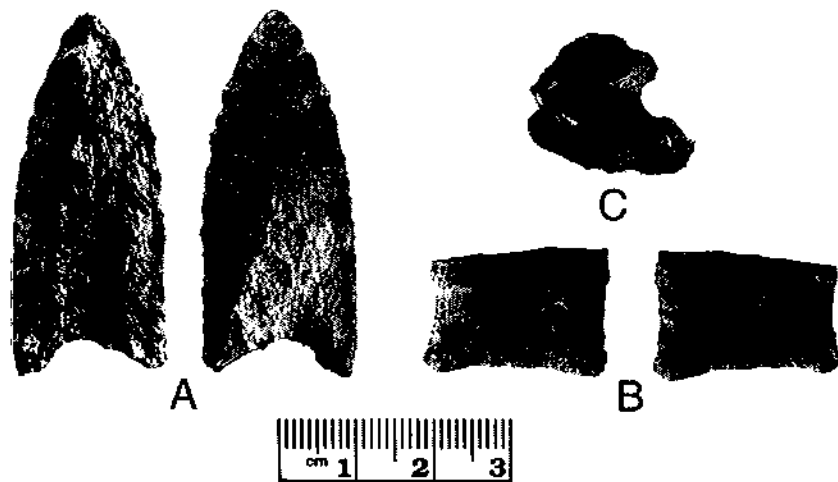
External contacts for the Israel River Complex are evidenced by the presence of exotic raw materials here and the recovery of flow banded rhyolites at other Paleoindian sites. Exotic raw materials occur in all of the Jefferson sites as fluted points, unifaces, retouched flakes and debitage. Pollock (1998) has defined criteria for six varieties of Munsungun chert, and based on inspection of the Jefferson tools collected through 1998 he has identified specimens that are consistent in visual criteria with Munsungun cherts. There are a total of 23 fluted points and fluted preforms in the assemblage. Seven are of Munsungun chert; two are of an unknown chert; and the others are flow-banded rhyolite from either Jefferson or Mt. Jasper. All the Munsungun specimens are points manufactured to completion, while five of the 14 rhyolite fluted pieces were so worked. Of the 16 intact end scrapers in the combined assemblages, three are of Munsungun chert. Only small amounts of exotic debitage are encountered, less than 2% of the estimated 20,000 waste flakes. Taken together, it appears that the tools made from exotic raw materials were entering the site as completed implements that were either resharpened on site or discarded as they came to the end of their use-life. These data indicate that the people had access to chert outcrops 180 km to the northeast in Maine and their tool kits were augmented by manufacture on site using local rhyolite. This suggests that points of local material were being made on site and exported. Recovery of fluted points at the Neponset site in southeastern Massachusetts made from flow-banded rhyolites traceable to northern New Hampshire supports this interpretation and extends the geographic network southward at least 320 km (Carty and Spiess, 1992; Bradley, 1998; Pollock, Hamilton, and Boisvert, 1998). Contacts to the west are inferred by the recovery of these same flow-banded rhyolites at the Fairfax Site in northern Vermont (Bradley, 1998:22). The Israel River Complex is lodged within a zone of travel and interaction that minimally encompasses New Hampshire, much of Maine, northern Vermont, and eastern Massachusetts.

Reliable radiocarbon dates are not yet available for the Jefferson Paleoindian sites. The best indicator for the periods of occupation is the stylistic variation of the projectile points. Spiess, Wilson and Bradley (1998) and Curran (1999) subdivide the era along roughly comparable lines, using variation in styles modeled after trends defined further west in Ontario and Michigan by Ellis and Deller (1988, 1997). Two fluted point variants are present in the Israel River Complex, an earlier Bull Brook-like or Gainey-like point, with essentially parallel sides and a relatively deep indented base, and a Barnes-like point with somewhat flared basal corners, a shallower indented base, and frequently with multiple fluting. Among the fluted points recovered from the Jefferson sites, five are assignable to the Gainey style and four to the Barnes style (see Figure). Also, an analysis of the manufacturing methods evident among the specimens broken in manufacture indicates a process consistent with Gainey manufacturing methods (Boisvert, 1999b). To date, we have recovered no Vail/Debert style points or the later unfluted Paleoindian points. The age and cultural affiliation estimate for the Israel River Complex is late Early Paleoindian and Middle Paleoindian.

In summary, the Israel River Complex is composed of at least five sites, each of which is exclusively Paleoindian in age and whose assemblages are dominated by locally obtainable flow-banded rhyolite. The sites contain multiple loci and it is likely that additional loci and sites remain to be discovered. Thus far 13 loci have been identified on the five recorded sites, but only one has been comprehensively excavated. Our understanding of the Complex is very limited and will certainly evolve as analysis proceeds and additional investigations are completed. Even so, some basic interpretations can be offered here. The Paleoindian presence in Jefferson reflects repeated occupations by hunting bands. Caribou on annual migrations through the Israel River valley could have been a prime prey. The lithic



Fluted points from the Israel River Complex. Top Row: Neponset/Barnes Style; Bottom Row: Bull Brook Style.



Jefferson IV Site. A: Complete fluted point, B: Basal fragment of a fluted point, C: Munsungun chert flake.

assemblages contain additional tools that would have been most appropriate for hide processing, clothing manufacture, and for making items from bone and antler. Abundant debitage and bifaces broken in manufacture attest to an active chipped stone tool production, at least part of which was geared toward replacing projectile points broken in use. The availability of the high-quality flow-banded rhyolite, exposed on the till surfaces and probably in the nearby brooks, provided a strong attraction to the specific site locations and may have been a factor in the larger settlement system. However the relative importance of this factor cannot yet be determined. Given these multiple tasks and their integration into the same loci, it is proposed that the sites functioned as extended-stay campsites for hunting bands that included entire family units and that these sites were reoccupied over several seasons. The import of finished tools made from exotic raw materials and the export of flow-banded rhyolites (some from the immediate area as well as from Mt. Jasper located only 30 km away) indicates that these people were part of a highly mobile population occupying a network of sites covering most of New England. The location of the sites on a major east-west thoroughfare, and relatively near to north-south routes along the Connecticut River or through the notches in the White Mountains, suggests that the Complex is a nexus within this network. The role of the Israel River Complex within this larger system appears to be distinct and unusual. It is clear that these sites are significant and that they hold great potential to add to our developing understanding of the Paleoindian culture in the Northeastern United States.

ROAD LOG

Assembly time and place: Saturday, May 18, 2002. 8:00 AM. Parking lot at Eastgate Motor Inn. The *New Hampshire Atlas*, published by DeLorme Mapping Co., is recommended for following the road log. An overview of the White Mountains is provided by the Mount Washington 1:100,000-scale topographic map. The directions and mileages given below will guide you from one stop to the next. Cumulative mileages are not given, because they will vary depending on driving distances in turnarounds, large pits, etc. **Note:** *Nearly all of the stops are located on private property. Permission must be obtained from the owners for any future visits!*

From the Eastgate Motor Inn, drive east on U.S. Rte. 302 to jct. with U.S. Rte. 3 in Twin Mountain. Turn L at traffic light onto Rte. 3 and drive N for 2.1 mi. Park on L, at entrance to Twin Mountain Sand & Gravel pit.

STOP 1: TWIN MOUNTAIN SAND & GRAVEL PIT AND ROCK QUARRY, Carroll

STOP 1-A: Overview of Carroll Delta and glacial Lake Ammonoosuc (W. Thompson)

The large pit complex at Stop 1 is located in an ice-contact delta known as “the Carroll delta” (Lougee, 1940), which built southward into glacial Lake Ammonoosuc. This deposit formed when the receding late Wisconsinan ice margin was pinned against Beech Hill to the west and Cherry Mountain to the east. The delta plain has an elevation of approximately 450 m (1475 ft) (Figure 4). It was graded to the Gale River 2 stage of glacial Lake Ammonoosuc, which had a spillway at about 445 m (1460 ft). This spillway can be seen along U. S. Route 3 southwest of Twin Mountain village (G2 in Figure 3).

Deposits in the vicinity of Twin Mountain record successively lower levels of Lake Ammonoosuc and the resultant downcutting of part of the Carroll delta. Figure 4 shows the relationship of the Carroll delta to meltwater channels, inferred ice-margin positions, and other features associated with the glacial lake. A cluster of end moraines located a short distance north of Beech Hill are also evident on Figure 3. Goldthwait (1916) described the Carroll delta as a “pitted outwash plain”, and proposed that it formed shortly before the Bethlehem Moraine. He noted the terraces extending south from the main part of the

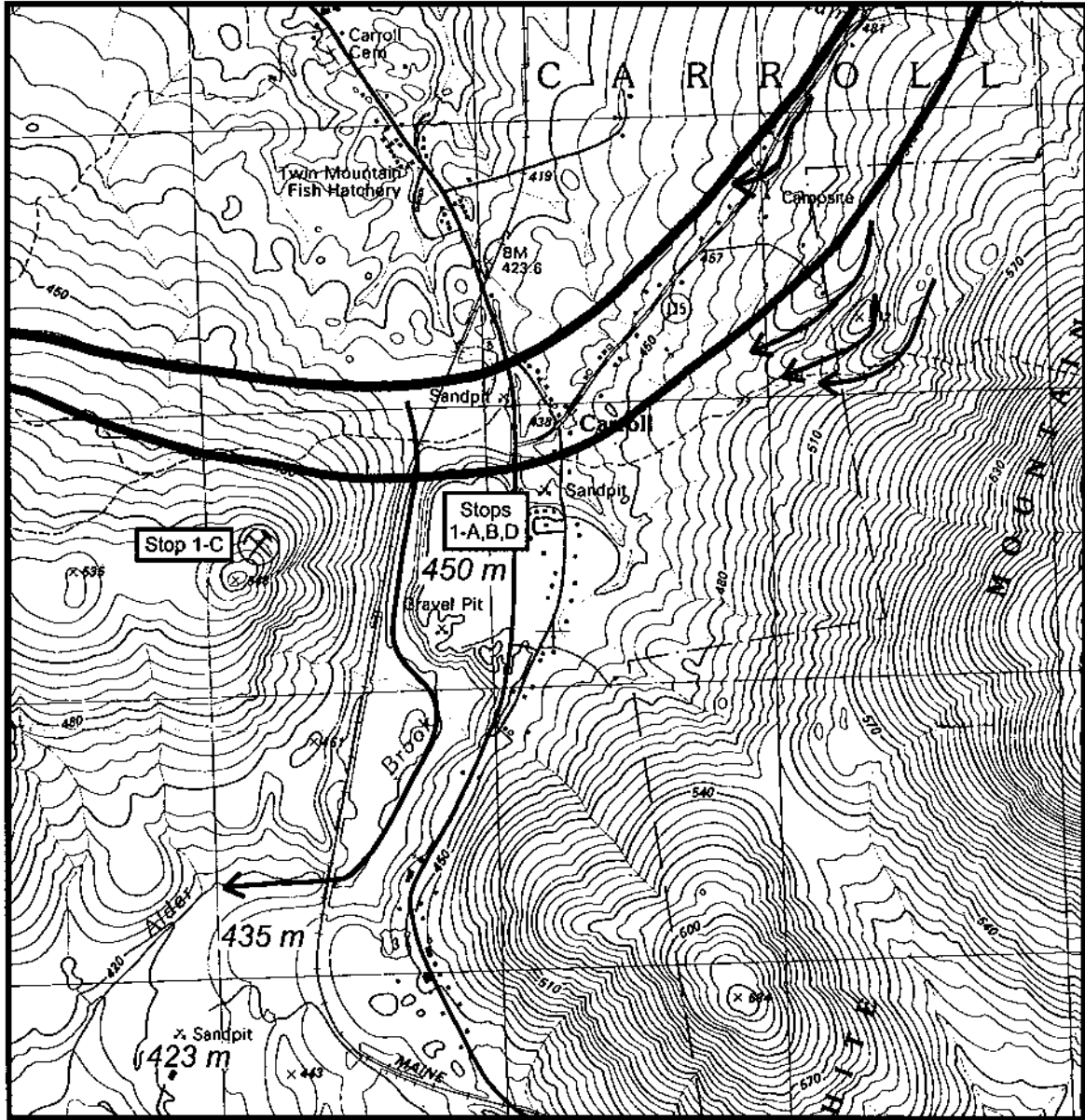


Figure 4. Part of the Bethlehem 1:25,000 metric quad, showing locations of Stops 1A-1D; the ice-contact Carroll Delta at elevation of ~ 450 m; the 435-m and 423-m surfaces graded to lower stages of glacial Lake Ammonoosuc; inferred ice-margin positions (gray lines); and meltwater channels (arrows). The long channel cutting through the delta is the Carroll Spillway. Modified from Thompson *et al.* (1996).

delta, along both sides of Alder Brook, and described them as kame terraces deposited against remnant ice in the Alder Brook valley.

Lougee (1940) examined this area in detail and agreed that the Carroll delta was deposited in Lake Ammonoosuc. However, he demonstrated that the Alder Brook terraces are not kame terraces, but simply resulted from downcutting of the distal part of the delta. Lougee (n.d.) understood that this incision resulted from a drop in lake level when recession of an ice tongue farther down the Ammonoosuc Valley opened up lower spillways (Figure 3). The deep channel that was cut into the west side of the delta as the lake fell is clearly seen along the railroad track at Stop 1. The elevation of the channel at this point is approximately 429 m (1407 ft). The water that discharged through it may have come initially from the northern ice-margin shown in Figure 4, and later from an ice-contact lake (glacial Lake Carroll) that formed as the ice margin receded northward from the Carroll delta. (Lougee proposed a third possibility, which is discussed below). The initial drainage of Lake Carroll through the channel would have been very short-lived, because ice retreat from Carroll village would have quickly opened a lower spillway at about 387 m (1269 ft) just northwest of Beech Hill (Figure 3).

Lougee (1940) described meltwater channels on the hillside northeast of the Carroll delta, which are evident in Figure 4. He thought that these channels resulted from the sudden drainage of a glacial lake that lay to the northeast of Cherry Mountain. Water escaping along the ice margin, around the north and west side of the mountain, supposedly carved these channels and deposited the eroded sediment onto the Carroll delta. Lougee said that the channels on the delta surface originate at the northeast corner of the delta, where the hillside drainage would be expected to have entered the delta plain. A problem with this scenario is that the channels can be traced only a short distance. They are absent from the remainder of the proposed drainage route along the flank of Cherry Mountain. Moreover, aerial photographs taken in 1955 (prior to the pit operation) show a network of subparallel channels trending from north to south across the full width of the delta plain. These photos also show a steep ice-contact slope on the north edge of the delta, where the pit is now located. Thus, even though meltwater and sediment probably issued from the short lateral channels on the hillside, it appears that the high-level (450 m) Carroll delta also received meltwater directly from the adjacent ice margin to the north.

There were no large fresh exposures in the pit when this guidebook was prepared. However, the coarse gravel forming the delta topset beds can be seen along the upper east wall of the pit, and sandy foreset beds and other interesting features are exposed in places (Stop 1-D). According to Goldthwait (1916), the delta plain (which has been removed in the pit area) was “strongly pitted” in the proximal part, but smooth and uncollapsed in the central and distal portions. He also observed that the topset gravel was less bouldery toward the front of the delta.

STOP 1-B: Coring and dating the spillway channel (C. Dorion)

A sediment core was obtained from the deep meltwater channel that cuts the western part of the Carroll Delta. This spillway channel was eroded by drainage from glacial ice to the north. Two days of probing were required in an effort to find a buried basin on the channel floor that would have accumulated organic material once the channel ceased to carry meltwater. We finally located and cored a suitable part of this spillway where sands on the channel floor contain typical late-glacial flora along with sedge, rush, and St. Johns Wort seeds. The wet channel environment provided the necessary conditions for these plants to exist. The basal radiocarbon age (see Table 1) shows that the vegetation grew on the channel margins by 11,430 BP. The channel itself most likely 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 glacial Lake Israel in the Cherry Pond area,

TABLE 1. LOCATION AND SIGNIFICANCE OF RADIOCARBON-DATED SITES IN THE NORTHERN WHITE MOUNTAINS (from Dorion, 2002)

Site name	Latitude (N)	Longitude (W)	Facies	Lab Accession Number	Reported Age (yr B.P.)	Material	$\delta^{13}C$ (‰)	Dated Event
Cherry Pond (1)	44 22 32.87	71 30 44.49	varves	PL-0000483A	7,760 ± 140	<i>Dryas integrifolia</i> , insect parts, <i>Picea</i> seed, <i>Cyperaceae</i> achenes, <i>Daphnia</i> sp.	-25.9	Deglaciation of Israel valley (minimum)
Cherry Pond (2)	44 22 32.87	71 30 44.49	varves	PL-0000496A	11,480 ± 80	insect parts, <i>Daphnia</i> sp., woody twigs, seeds	-23.6	Drainage of Glacial Lake Israel
Cherry Pond (3)	44 22 32.87	71 30 44.49	varves	PL-0000489A	11,800 ± 80	<i>Dryas integrifolia</i> leaves, insect parts, <i>Daphnia</i> sp., <i>Vaccinium oxycoccos</i> leaf, <i>Betula</i> bract	-22.9	Drainage of Glacial Lake Israel
Cherry Pond (8)	44 22 32.87	71 30 44.49	gyttja	PL-0000485A	10,790 ± 80	insect parts, <i>Daphnia</i> sp., <i>Picea</i> needles	-26.9	Termination of Bolling-Allerod
Cherry Pond (4)	44 22 32.87	71 30 44.49	lithic zone	PL-0000488A	10,720 ± 80	insect parts, <i>Daphnia</i> sp., <i>Betula</i> bract	N.D.*	Onset of Younger Dryas lithic zone
Cherry Pond (6)	44 22 32.87	71 30 44.49	lithic zone	PL-0000486A	10,370 ± 80	insect parts, <i>Daphnia</i> sp., <i>Picea</i> seeds and needles, <i>Larix</i> needle, woody twigs, <i>Betula</i> bracts, <i>Salix</i> twig, <i>Vaccinium uliginosum</i> leaves	-27.2	Younger Dryas lithic zone termination
Cherry Pond (7)	44 22 32.87	71 30 44.49	gyttja	PL-0000484A	10,070 ± 80	insect parts, <i>Picea</i> needles, woody twigs, <i>Betula</i> bracts, <i>Vaccinium uliginosum</i> leaves, <i>Carex</i> sp. achenes	-27.1	Onset of Holocene
Cherry Pond (5)	44 22 32.87	71 30 44.49	gyttja	PL-0000487A	8,440 ± 80	gyttja	-24.8	Peak organic productivity in lake
York Pond (12)	44 30 10.09	71 20 14.08	varves	PL-0000494A	11,980 ± 90	insect parts, <i>Daphnia</i> sp., <i>Picea</i> needles and seeds, woody twigs, <i>Vaccinium uliginosum</i> leaves, <i>Carex</i> sp. achenes	-28.0	Deglaciation of York Pond
York Pond (13)	44 30 10.09	71 20 14.08	gray mud	PL-0000495A	10,900 ± 80	insect parts, <i>Dryas integrifolia</i> leaves, woody twigs	-25.6	Bolling-Allerod time
York Pond (10)	44 30 10.09	71 20 14.08	lithic zone	PL-0000490A	10,650 ± 80	insect parts, <i>Picea</i> needles and seeds, woody twigs, <i>Cyperaceae</i> achene	-27.5	Onset of Younger Dryas lithic zone
York Pond (11)	44 30 10.09	71 20 14.08	lithic zone	PL-0000493A	10,090 ± 80	insect parts, <i>Picea</i> needles and seeds, woody twigs, <i>Vaccinium</i> sp. leaves, <i>Betula</i> seed	-26.6	Younger Dryas lithic zone termination
York Pond (9)	44 30 10.09	71 20 14.08	gyttja	PL-0000492A	8,180 ± 80	<i>Picea</i> needles and seeds, <i>Pinus strobus</i> needles, <i>Betula</i> seeds, conifer bracts	-26.3	Peak organic productivity in lake
Martin Meadow Pd. (14)	44 26 29.03	71 36 12.62	varves	PL-0000491A	12,360 ± 80	insect parts, <i>Daphnia</i> sp., <i>Dryas integrifolia</i> leaves, woody twigs, <i>Salix herbacea</i> leaves,	N.D.	Deglaciation of Martin Meadow Pond
Martin Meadow Pd. (15)	44 26 29.03	71 36 12.62	varves	PL-0000889A	10,920 ± 80	insect parts, <i>Daphnia</i> sp., <i>Dryas integrifolia</i> leaves, conifer needles, woody twigs, <i>Cyperaceae</i> seed, <i>Brassica schreberi</i> seed.	-26.0	Deglaciation of Martin Meadow Pond
Carroll spillway (16)	44 17 23.82	71 32 46.13	glaciofluvial sand	PL-0000890A	11,430 ± 80	insect parts, wood pieces, <i>Dryas integrifolia</i> leaves, <i>Eleocharis</i> and <i>Scirpus</i> seeds, <i>Cyperaceae</i> seeds, <i>Hypocnemum</i> seed, <i>Juncaceae</i> seeds	-26.0	Carroll spillway termination

*Not Determined

where we have ages as old as 11,800 BP. The cores that we recently obtained from Cherry Pond and two other ponds in the northern White Mountains are discussed under Stop 3.

STOP 1-C: Sub-till saprolite at Beech Hill quarry (B. Fowler / W. Thompson)

A rock quarry was opened on the east end of Beech Hill to supply a crushed stone operation in the gravel pit. Lyons *et al.* (1997) mapped the bedrock in this area as foliated "biotite granite" belonging to the Oliverian plutonic suite of Ordovician age. However, a variety of rock types including quartzite and metavolcanics are evident in the quarry face (Timothy Allen, pers. comm.).

Just west of the quarry, near the top of the hill, a thin cover of glacial till has been scraped off the ledge to prepare for expansion of the quarry pit. Two small sections will be examined on the edge of this scraped area. These sections are still being studied, and the following descriptions are very tentative. A 2-m section shows a thin (~0.3 m) till unit overlying an unusual diamict. The till is compact, fissile, and pebbly, with a light olive-gray to brownish gray (oxidized) silty-sandy matrix. It contains much granitic rock like that seen in the quarry, but also includes several other rock types. Some of these clasts are faceted and striated, and they are generally fresh. The underlying diamict unit contains abundant larger stones, many of which are more-or-less rotten. The rock types in the diamict include dark-green chlorite schist (Ammonoosuc volcanics), black amphibolite schist, and granitic lithologies. The diamict also contains steeply to vertically-dipping sandy layers of various colors. At first these were thought to be glacial sand lenses, but closer inspection reveals that at least some of them are disintegrated rock. The layered structure may have resulted from glaciotectionic crushing and deformation of rotten clasts. Just below the contact with the till, the rottenstone layers have been deformed by drag at the base of the overriding ice and smeared out parallel to the contact. The till contact dips 10° to the WSW, and the sense of drag along the contact is generally toward the east. The latter orientation suggests eastward glacial flow, but this direction of ice movement is anomalous when compared to the regional SE to S ice-flow directions recorded by most striations.

The compaction and fissility of the upper unit in this section, and its structural relation to the diamict, suggest it is lodgement till. The diamict appears to consist mostly of deformed rottenstone that may have been mixed with some fresher glacial clasts. The rottenstone material probably was transported only a short distance from its source; otherwise it would have been thoroughly dispersed and mixed with other glacial sediment to form a more typical till.

The other section is on the north edge of the scraped area. It is about 3 m high and shows 1 m (remaining below scraped surface) of lodgement till sharply overlying rotten, fine, sugary, gray schist/granofels with lenses of disintegrated white granitic rock. The till is oxidized to a brownish color and shows good fissility in the upper part. It is uncertain whether the rottenstone is an *in situ* saprolite or if it has been glacially transported. However, it appears less disturbed than the chaotic diamict in the section described above. It is unusual to find a remnant of saprolite in such an exposed location as the top of Beech Hill. The sections at this site will be discussed in relation to other New Hampshire sub-till saprolites described by Goldthwait and Kruger (1938). These authors reported a locality somewhere near here, in the town of Carroll, where they observed "till full of rotted blocks, and perhaps rotted ledge" in an area likewise underlain by granite gneiss bedrock.

STOP 1-D: Section in proximal part of Carroll Delta, and discussion of Lake Ammonoosuc spillways (C. Hildreth)

This delta is the ice-contact head of outwash for a major part of the deposits laid down in the Gale River 2 Stage of glacial Lake Ammonoosuc, whose outlet was near the south end of the Bethlehem East quadrangle, along Route 3 at elevation 447 meters (1460 ft.). The present excavation in the proximal part of the Carroll Delta exposes a diminished version of a large fresh section observed in this pit face during the summer of 2000. (Two collages of year-2000 photos will be presented and described during the pit visit). At that time, the pit face exposed a vertical section totaling 6-15 m and extended from ESE to N over a distance of more than 120 m. The eastern two-thirds of the pit face exposed 1-2 m of undisturbed planar and horizontally-bedded gravel overlain by 1-1.5 m of undisturbed to barely disturbed, nearly horizontal, ripple-drift sand and silt beds, apparently draped over a very gently dipping broad subaqueous fan that has a roughly N-S axis. These materials were overlain by a 3-6 m package of sand and silt beds that were variably disturbed by overturned folds and low-angle thrust faults that strike roughly NE and dip NW. For the most part, relative motion was up toward the SE. The amount of displacement is difficult to assess, but appears to be minimal (apparently less than 2 m along each individual fault), though many appear to become bedding-plane faults when traced any distance. The present-day exposure is largely concealed by recent slumping, but in the central and eastern parts of the face, bedding deformed by folds and very low-angle thrust faults can be seen. Some of the faults strike NW and dip NE.

Toward the center of the pit face exposed in 2000, the package of disturbed sediments was overlain by 1-2 m of pebble gravel. This unit was horizontal in the center of the pit, but dipped north and became less well sorted where the pit face began to turn northward; it also cut down across the disturbed package of silt and sand beds that comprised much of the pit face to the east. Near the bottom of the exposure of this gravel unit, where it dipped northward below slumped material, it was underlain by horizontally bedded, well-sorted gravel similar to that seen at the bottom of the pit at the east end. Also, where the pebble gravel unit began to dip north, it was overlain by 1-2 m of very boulder-rich diamict, which in turn graded upward into 1-3 m of conformable, less boulder-rich diamict similar to the common till of the area. The uppermost 1 m of this unit was apparently crudely layered parallel to the N-dipping top surface of the unit. In the northernmost part of the pit face, this unit was unconformably overlain by 1-3 m of horizontally bedded sand and silt layers interpreted to be bottom or outwash sediments of the earliest stage of glacial Lake Carroll, which occupied the valley to the north and whose first outlet was through the Carroll Delta following retreat of the ice from the ice-contact delta head.

The disturbed bedding and overlying north-dipping gravel and till are interpreted as ice-front readvance features that may be related to the Littleton-Bethlehem Readvance. However, these features may be too far south to be correlated with the main part of that episode, and probably represent only a short readvance somewhat earlier than the Bethlehem event.

The lowermost deposits uncovered at this site during the 2000 excavation were dug out below the present pit floor in a narrow N-trending 3 m-deep trench. The backhoe operator was "chasing" gravel, which turned out to be well-sorted cobble gravel. This deep gravel is interpreted to be part of the esker feeder that is related to the exposed esker segments in the Ammonoosuc River valley south and southeast of here. The esker system graded to Crawford Notch and the Crawford Stage deposits of glacial Lake Ammonoosuc.

When the outlet of the Gale River 2 Stage of glacial Lake Ammonoosuc was abandoned for the lower Stages 3 and 4 outlets (435 m [1427 ft] and 423 m [1387 ft] respectively), south-flowing meltwater

cut through the Carroll Delta, eroding the older deltaic sediments along its course and redepositing them farther south into the lowered lake. Some of the coarsest material consisting of pebble-cobble gravel and sand and gravel was deposited in lower deltas in glacial Lake Ammonoosuc that have surfaces of about 432 m and 420 m, north of Route 302. Pits in these lower delta deposits expose clast-supported pebble-cobble gravel. The orientation of many of the asymmetric clasts in these deposits indicate a SW current direction (Close-up photos of these gravel clasts will be shown at this stop).

Outlet channels of Gale River Stages 2 – 5 of glacial Lake Ammonoosuc. Drainage from the Gale River 2 Stage of glacial Lake Ammonoosuc flowed SW through an outlet channel near the southwest edge of the Bethlehem East (BEQ) quadrangle (elevation 447 m [1460 ft]). This flat-floored, swampy channel follows first the north, and then the south side of Route 3 for about 2 km SW. Then it bends northward in the South Twin Mountain quadrangle and reenters the BEQ, forming the main channel of Beaver Brook. This brook is an underfit stream which flows briefly N, thence W into the Bethlehem West quadrangle, thence S into the Franconia quadrangle, joining the North Branch of the Gale River at the point where the original dam for glacial Lake Gale is inferred to have stood. Today, outlet channels for Gale River Stages 3-5 form tributaries of the Beaver Branch which join the main stem of that branch at the SW corner of Bethlehem East and SE corner of the Bethlehem West quadrangles.

Beaver Branch Stage deposits of glacial Lake Gale. Deposits of the Beaver Branch Stage of glacial Lake Gale (Hildreth, 2000) are graded south and west to progressively lower spillways cut into till, between 424 and 402 m (1400 and 1320 ft) altitude. These spillways are located along a narrow stretch of the Gale River valley between Abbot Hill on the south and the east flank of Cleveland Mountain on the north, along Route 3 in Bethlehem, about 0.3 km south of the border between the Bethlehem East and Franconia quadrangles. The landfill site near the south edge of the Bethlehem quadrangles exposed 1.2 m of pebble-cobble gravel overlying 4 m of interbedded clay, silt, and very fine sand with scattered waterlaid till or flowtill lenses (Hildreth, 1984). Recent exposures in the same area revealed 3.7 m of alternating compact silt to very fine sand and flowtill layers, each of which is less than 1 m thick. All these materials indicate the proximity of the ice front during deposition. Some of the Gale River Stages of glacial Lake Ammonoosuc drained across these earlier deposits of glacial Lake Gale, which apparently filled with sediment as its outlet was progressively lowered by erosion. The spillways of the last four Gale River stages (2-5) of glacial Lake Ammonoosuc incise the Lake Gale Beaver Branch Stage sediments. Three stages of Lake Gale are differentiated in this area, based on topography, presence of flowtill and deformed sediments, and relationship to outlet channels of the Gale River stages of glacial Lake Ammonoosuc.

Continue N on Rte. 3 for 2.3 mi. Take sharp L onto gravel rd. Drive through field and up onto northern spur of Beech Hill, keeping L at first fork. Go 0.9 mi from Rte. 3 and park adjacent to small pit on R side of rd.

STOP 2-A: BEECH HILL TILL PIT, Carroll (C. Hildreth)

At this locality we will examine a new pit exposure of late Wisconsinan till which contains abundant grüss. This granitic debris presumably was derived from weathered and disintegrated local bedrock of the Oliverian Plutonic Suite (Lyons *et al.*, 1997).

Continue SSW on woods rd. for 0.5 mi. Park in cleared area on either side of rd.

STOP 2-B: BEECH HILL MORAINES AND MELTWATER CHANNELS (W. Thompson)

In 1997, W. B. Thompson checked a cluster of ENE-trending ridges seen on the lower north side of Beech Hill on the Whitefield 15-minute topographic map (Figure 3). The contour pattern on the map suggested these ridges might be moraines like those in the Littleton-Bethlehem area. This interpretation proved correct, and the Beech Hill moraines were correlated with the Bethlehem Moraine complex to the west and a series of moraines in the Israel River valley to the northeast. Thompson *et al.* (1999) attributed all of these moraine fields to the Littleton-Bethlehem Readvance, whose southern limit is shown in Figure 1. Quadrangle mapping by Hildreth (2000) has revealed the presence of many other moraines in the area between this stop and Whitefield village. Recent logging operations in the vicinity of Stop 2-B removed the tree cover from the Beech Hill moraines and made them much easier to see.

The Beech Hill moraines occur at elevations of 396-427 m (1300-1400 ft). They are till ridges which are 4-14 m high, up to 700 m long, and trend ENE-WSW. An exposure in a small borrow pit shows light olive-gray till that is loose, sandy, stony, and non-stratified. Large boulders of coarse white granite (up to 3 m wide) are extremely abundant on the surfaces of the moraines. They were derived from the local bedrock, which is an extensive body of "moderately to weakly foliated biotite granite" belonging to the Oliverian Plutonic Suite (Lyons *et al.*, 1997). Many of the granite boulders are weathered and more-or-less disintegrated. Some contain xenoliths, and they may be cut by mafic or aplitic dikes. A greater variety of rock types were noted in the till pit, including igneous and metamorphic lithologies. The non-granitic stones are generally small and would not be so easily spotted on the vegetated moraine surfaces. Some of these till stones are rounded or striated.

Meltwater channels occur on the distal sides of the Beech Hill moraines. The largest channel lies along the southern margin of this moraine cluster. It hosts a wetland and small pond seen on the newer Bethlehem quadrangle map. The channels developed sequentially in the swales between moraine ridges and thus are parallel to the moraines. Meltwater flowed southwest along the ice margin as it retreated from Beech Hill and drained through the 387-m gap between the northwest corner of Beech Hill and Pine Knob (Figure 3).

Return to Rte. 3 and turn R. Drive 0.7 mi. and turn L onto Lennon Rd. Continue 1.8 mi. to Rte. 115 and turn L. Go 1.9 mi. N on Rte. 115 and turn L into scenic turnout.

STOP 3: CHERRY POND OVERLOOK AND RESULTS FROM LAKE SEDIMENT CORES, Jefferson (C. Dorion)

From this vantage point, there is a good view of the Johns River basin (foreground). To the right, we look north across Cherry Pond to the Israel River valley and the upper Connecticut River valley in the far distance. Cherry Pond is located either just within the limits of glacial Lake Israel (Baileys Stage), or in the headward part of the spillway that drained Lake Israel waters southwest into glacial Lake Whitefield. The elevation of this spillway is about 338 m. Lake Whitefield occupied the Johns River valley, and the Whitefield airport is situated on a delta built into Lake Whitefield.

In 1999 we obtained three cores (A, B, and C) from the basal Cherry Pond sediments. The varves in the lower part of these cores were deposited into glacial lake Israel. We attempted to date the earliest varves, which presumably were deposited at or near the glacier margin. This would tell us when the valley became ice-free and provide a limiting date for earliest possible occupation by Paleoindians. However, after careful sieving of the A, B, and C cores, we were disappointed to find only a few degraded pieces of organic material in the bottom varves. From other pond cores in the area, we know

that vegetation, insects, and other fauna were present at the glacier margin. The margin was receding very slowly, approximately 50 to 150 m per year on average, during late-glacial time. Our hypothesis as to why so little organic matter became incorporated into the varves relates to how glacial meltwater discharged at the ice sheet margin.

In one scenario, glacial meltwater exited a tunnel mouth at the ice margin, flowed downslope as a stream, and then entered the Lake Israel. Vegetation was periodically washed into the meltwater stream and carried out into the lake where it settled to the bottom along with the clay, silt, and sand to form a varve. However, in the Israel valley glacial meltwater discharged through a tunnel at the base of the ice margin and directly into the lake. Since the glacial stream dumped its sediment load directly from the ice tunnel into the lake, it could not incorporate vegetation living on the nearby hills. Perhaps small nonglacial streams could have carried organic material into the lake? This is not likely as there is abundant data supporting an arid steppe environment during late-glacial time; the streams so prevalent today in the White Mountains would have been dry most of the year. In Greenland today, along the ice sheet margin, only about 15 cm of annual precipitation falls, mostly as snow. The White Mountains average over 100 cm today with significantly higher amounts at higher elevations.

The 7,760 yr BP radiocarbon age from the Cherry Pond cores (Table 1) was obtained on a composite sample of very sparse organics from the basal portions of all three cores. Comparison with the other results from Cherry Pond shows that this age is clearly anomalous. We also obtained two radiocarbon ages from the youngest varve in the cores (Table 1). These ages suggest that the ice sheet receded from the central Israel valley by 11,800 ^{14}C yr BP. We can estimate the time of deglaciation because Jack Ridge (Tufts University), using a modified gray-scale imaging system, counted 90 varves from the base of the varved section of the core to the top, suggesting that glacial meltwater entered the lake for 90 years before being diverted elsewhere. This would place the time of deglaciation close to 12,000 BP, in close agreement with the other evidence we have gathered.

For the next 1,000 years following the demise of glacial Lake Israel, the bottom sediments of Cherry Pond were mixed to some extent by burrowing organisms. Due to the arid conditions, the lake was most likely alkaline and supported aquatic gastropods. It probably was smaller than today and hence shallower. The high silt and clay content of this section of the core reflects an open landscape with bare ground in places and wind-deposited loess (silt and fine sand). The landscape was similar to that of the Great Plains of the central U.S. with trees only occurring along ponds and other wet areas. Mean annual temperatures were significantly lower than today. Most likely, the climatic extremes we now experience did not exist. Periodically, high winds would entrain silt and fine sand from glacial outwash and river alluvium in valleys and deposit this loess as a blanket across the landscape. An ongoing research question regarding the silt and fine sand sections we find in pond cores during this time is: To what extent was the sediment of eolian (loess) origin? Or, was the sediment deposited by water eroding the landscape?

Beginning at about 10,700 BP, an abrupt climatic cold period began, called the Younger Dryas chronozone. Its onset was rapid as seen in the razor-sharp transition in the core from silty organic-rich gyttja back to barren gray mud. Records from the Greenland Ice Sheet show the onset occurred in 3 to 20 years (Stuiver *et al.*, 1995). Global temperatures dropped abruptly and the major ice sheets of the northern Hemisphere as well as glaciers in the southern Hemisphere expanded once again (Lowell *et al.*, 1995). It was during this abrupt climatic change that Paleoindians moved into northern New England. We do not see any sedimentary evidence in the Cherry Pond record that glaciers advanced into the Israel valley in Younger Dryas time.

The Younger Dryas chronozone terminated as abruptly as it started, and the Holocene Epoch began at about 10,070 BP. This is noted in the cores as a return to the organic-rich gyttja. By about 9,500 BP, the Paleoindian tradition had vanished in northern New England. The once-open steppe was quickly closing in with the forests we know today as mean annual precipitation dramatically increased. Where foot travel was once rapid and visibility nearly unlimited, dense forest now covered New England, restricting travel greatly.

The cores and radiocarbon ages that we obtained from York Pond in Berlin show stratigraphy and chronology similar to Cherry Pond (table 1). This record confirms most of our previous interpretations in the area. York Pond is a kettle pond located on a valley wall composed of abundant glacial deposits such as eskers, fans, and end moraines. The base of the York Pond core is likewise composed of varves, from which the basal organic material was sieved. The varve section is much shorter than at Cherry Pond; the topographic setting of York Pond did not favor a prolonged glacial lake stage.

It is important to note that the radiocarbon ages from Cherry and York Ponds that bracket the onset and termination of the Younger Dryas climatic cooling event are in close agreement with records from Maine (Dorion, 1997) and maritime Canada (Stea and Mott, 1989). Also, the ages coincide with those from nearby Pond of Safety in Randolph, which we described for the 1996 NEIGC trip (Thompson *et al.*, 1996). Together, these new data show that the cooling event was abrupt and had an immediate effect upon the landscape across northeastern North America. The Paleoindians witnessed this dramatic shift in climate in the northern White Mountains and most likely witnessed its similarly abrupt termination nearly a millennium later.

The two basal ages from Martin Meadow Pond in Lancaster (12,360 and 10,920 BP, Table 1) show a wide spread. However, the younger sample was taken from a subsidiary basin of the pond. It is likely that during late-glacial time, some mixing of younger sediments occurred in the area where we took this set of cores. In general, the oldest age from a particular pond basin is the one most likely to reflect the true time of deglaciation.

Continue NE on Rte. 115. About 1.9 mi. from overlook, note historic site marker on R for the **Cherry Mountain landslide (1885)**. Proceed 3.9 mi. and turn R onto U.S. Rte. 2. Go 0.4 mi uphill on Rte. 2 and note one of the **Boy Mtn. meltwater channels** crossing the road. Other channels occur farther up the mountain and nearby on both sides of the Israel River valley. Continue 2.7 mi. E on Rte. 2 and make a sharp R turn onto Valley Rd. (This jct. is on a **Bowman Stage delta of Glacial Lake Israel**). Go 0.3 mi. W on Valley Rd. and turn L onto gravel access rd. leading a short distance into the Corrigan Pit.

STOP 4: CORRIGAN PIT, Randolph (W. Thompson)

Return to Rte. 2 and turn L toward Jefferson (*watch for traffic here!*). Drive 6.9 mi. on Rte. 2 to Jefferson village. Turn L onto Rte. 115A. Drive S for 1.3 mi. and turn R onto driveway/woods rd. leading to the next stop.

The Corrigan Pit (Figure 5) is located next to the Israel River, which flows northwest and joins the Connecticut River at Lancaster. A short distance east of here, at Bowman, Route 2 crosses the 457-m divide into the Androscoggin River basin. As the late Wisconsinan glacier margin receded down the Israel Valley, the Bowman Stage of glacial Lake Israel was impounded in this area and spilled eastward across the divide. Glaciolacustrine fans and deltas deposited in this lake have been exposed in several pits in Randolph and Jefferson. An ice-contact delta graded to the Bowman Stage forms a narrow terrace along Route 2 just north of the Corrigan Pit.

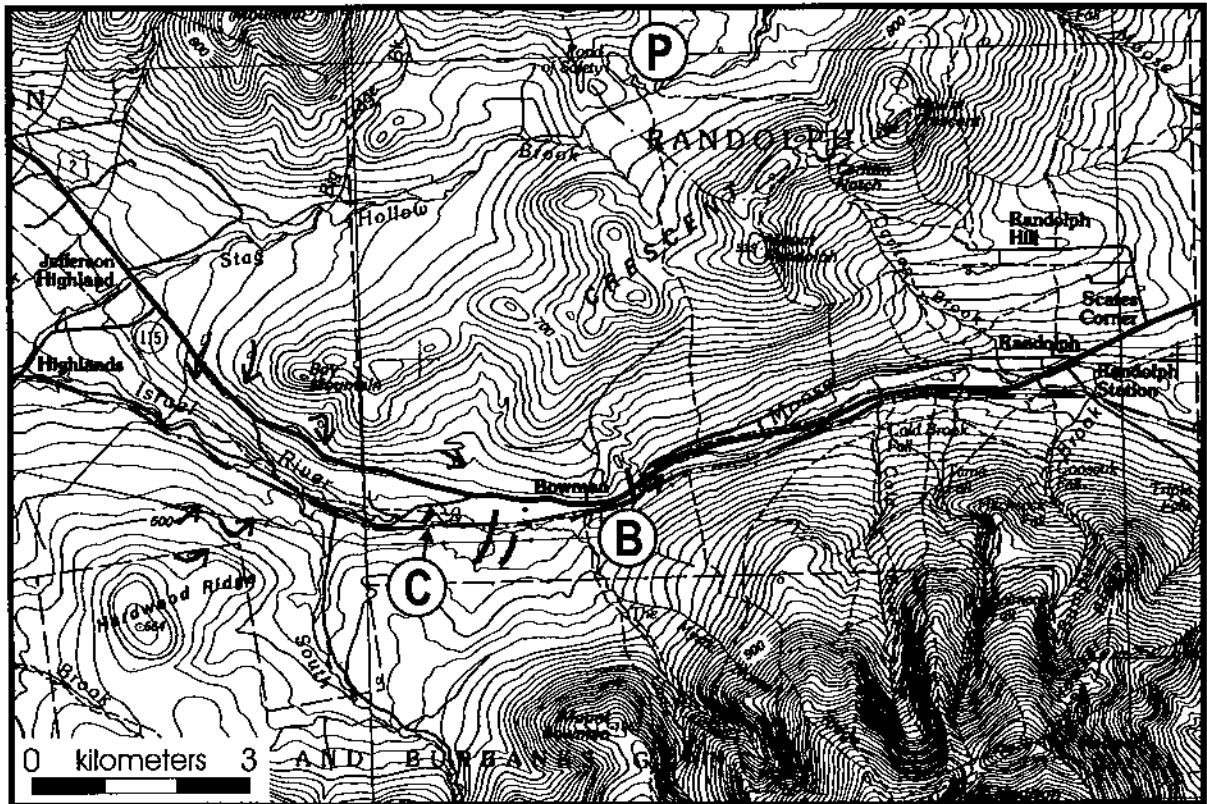


Figure 5. Map showing upper Israel River valley and sites mentioned in text. B: Bowman spillway for glacial Lake Israel. C: Corrigan Pit (Stop 4). P: Pond of Safety coring site, which yielded 12,450 BP radiocarbon age from basal pond sediments and stratigraphic evidence of Younger Dryas climate event (C. C. Dorion, in Thompson *et al.*, 1996). Arrows indicate meltwater channels. Heavy black lines near Bowman are moraines. From Thompson *et al.* (1999).

Northwest of here, meltwater channels were cut into the west side of Boy Mountain along the margin of the retreating Israel Valley ice tongue (Lougee, n.d., 1939; Goldthwait and Mickelson, 1982). Lateral meltwater channels and ice-contact deposits also exist on the south side of the valley (Crosby, 1934; Goldthwait and Mickelson, 1982; Thompson *et al.*, 1999). Lougee (n.d.) noted esker and lake-bottom sediments downvalley from here.

As we consider the glacial features in the Corrigan Pit and surrounding area, the principal question concerns the mode of deglaciation. By analogy with Alaskan glaciers, Goldthwait and Mickelson (1982) interpreted this area as demonstrating the early stagnation of the late Wisconsinan ice sheet, with development of a ragged ice margin as the glacier thinned over high-relief terrain. However, Thompson *et al.* (1999) concluded that the receding ice remained active. It built several prominent end moraines in the upper portion of the Israel valley and numerous small moraine ridges near Jefferson village.

The Corrigan Pit is located in a low moraine ridge with a surface elevation of up to 439 m. The sections in the southern part of the pit, close to the river, formerly exposed a sequence consisting of glaciolacustrine sand and gravel which coarsened upward and was overlain by stony glacial diamict. Much glaciotectonic deformation was evident in the sand and gravel. The principal stratigraphic units observed in the Corrigan Pit in the early 1990's are as follows: (1) The lowest unit consists of 2+ m of glaciolacustrine sand (base not exposed) with planar foreset beds dipping NE to E. In places there is much faulting in the central to upper parts of this unit, which probably resulted from ice shove. (2) The lacustrine sand is overlain by a variable thickness (~ 2-4 m) of ice-proximal sand and gravel. The gravel fraction is poorly sorted and angular. This unit contains scattered outside boulders and diamict lenses (flowtills). Local shear structures indicate ice shove to the northeast. (3) The uppermost unit consists of up to 6 m of bouldery, silty-sandy glacial diamict containing abundant lenses of washed sediment. This unit appears identical with the regional late Wisconsinan surface till. Some of the sand lenses in the diamict are highly deformed and suggestive of ice shove.

The deposits in the Corrigan Pit are believed to record the eastward advance of a glacier margin into a lake ponded in the Israel River valley. The constructional morainic topography suggests that these deposits resulted from a readvance into the Bowman stage of glacial Lake Israel during the overall recession of the ice sheet.

STOP 5: JEFFERSON II PALEOINDIAN SITE, Jefferson (R. Boisvert)

This site was the subject of excavations by the NH State Conservation and Rescue Archaeology Program in 1997 and 1998 under the direction of Dr. Richard Boisvert of the NH Division of Historical Resources.

The Jefferson II Site is large, encompassing a field of approximately one hectare and extending an undetermined distance into second-growth forests to the west and north. The original boundaries of the site cannot be determined because sand and gravel quarrying have impacted the southern portion of the site. The open field portion has been tilled, leaving a 15 to 20-cm deep plow zone. Most of the site is on sandy, stony glacial ablation till with abundant boulders. The till contains a high-quality rhyolite, typically flow-banded and spherulitic, which was utilized for the manufacture of tools on-site, including fluted points. The weathered upper part of the till is typically fine sandy loam with areas of intensely concreted soil, usually in association with spodic soil anomalies.

A single radiocarbon sample has been processed from Jefferson II but has been of little help in determining the age of the site. It was obtained from 46 cm below the surface in the till deposits in the large excavation block. It was associated with four very small flakes of what appears to be high-quality translucent white chert or chalcedony. The AMS date (Beta-108465) is 8,570 +/- 60 radiocarbon years. The calibrated date is 9,600 yr B.P., frustratingly short of the anticipated range of 10-11 ka. The date, while acceptable as a Late Paleoindian date, is not consistent with age determinations for fluted points, and we would estimate that it is 500 to 1000 years too young to be affiliated with the occupants who made and used the fluted points found at the site. The styles of the tools found at the site, especially the fluted projectile points, place the occupation at approximately 11,000 to 12,000 calendar years ago.

The area of heaviest artifact concentration is located within the till. A 43 square meter block was excavated, centered on the shovel test pit that produced the biface fragment. Excavations recovered 6 fragments of fluted points, as well as channel flakes, biface fragments, end scrapers, side scrapers, retouched flakes, and over 12,000 waste flakes. At least 3 other similar areas of occupational concentration have been documented and more are likely to exist. Based upon the multiple occupations at the site, and the patterned distribution of artifacts within each, we conclude that the site was used as a temporary base camp by a family bands. Caribou hunting was most likely the primary subsistence activity at the site, although it is almost certain that other hunting and foraging activities were also conducted.

Contacts with other regions in the Northeast are reflected in the sources of stone used to manufacture the various tools. Cherts from Munsungun Lake, Maine are represented in the tool inventory. Dusky red, pale green and gray/black banded varieties are represented. In addition, the locally occurring spherulitic banded rhyolite also appears at sites in southern Maine and Massachusetts. This supports the widely held interpretation that the Paleoindian peoples in the Northeast were engaged in long-range annual migration cycles and exchange with similar hunting bands.

The Jefferson II site is considered highly significant, and in 1998 the Archaeological Conservancy purchased the site in order to preserve it.

Return to Rte. 115A, turn L, and drive about 0.5 mi. back toward Jefferson. Park carefully as directed, on roadside or adjacent properties.. Site is across rd. from Apple Brook Inn.

STOP 6: JEFFERSON IV AND V PALEOINDIAN SITES, Jefferson (R. Boisvert)

In June 2000, a chance discovery of the base of a fluted point by the landowner marked the identification of the Jefferson IV in the Israel River Complex. In the following July and October, follow-up excavations by SCRAP recovered an additional (and unbroken) fluted point and related manufacturing debris. As all the other sites in the Complex, no artifacts from any later prehistoric era were recovered, and the site appears to have been occupied only during the Paleoindian period.

Although the artifact count and density are low, the site has yielded significant information. The fluted points represent two distinctively different styles. The complete point (see Figure) is most similar to the Gainey and Bull Brook varieties exhibiting ground parallel sides and a base indented 5 mm. The basal fragment most closely resembles the Barnes or Neponset style with slightly flared basal corners and shallow concave base. Both points were found in undisturbed strata and are the only diagnostic artifacts recovered from the site.

A Munsungun chert flake (see Figure) was discovered *in situ* in undisturbed context. Upon recovery we recognized that the ridges on the flake exterior were extremely worn, indicating the flake had been struck from a relatively large biface that had been used strenuously for cutting or scraping.

Also, the flake had a concavity created by the hinge termination of a previous flake removal and organic material would have been trapped in this concavity. Consequently we considered the specimen to be a prime candidate for the retention of animal protein. Residues removed from the flake and the complete point were subjected to a crossover immunoelectrophoresis (CIEP) analysis by Kathryn Puseman at Paleo Research Laboratories in Denver, Colorado, and tested against antisera from 18 different animal species. The complete point yielded negative results to all antisera tested. The flake, however, yielded a positive result to deer antiserum, indicating the presence of protein from a member of the Cervid family such as deer, moose, elk or caribou. Given the context of the find, we feel that caribou is the most likely source, although the other possibilities cannot be dismissed. This identification is important because there is scant evidence for specific fauna from New England Paleoindian sites

Concurrent with the October 2000 excavations on the Jefferson IV site, excavations were initiated on a nearby knoll on the property of the same landowner. Investigations there were quickly rewarded with the recovery of chipping debris, including flakes made from Munsungun chert. In September and October 2001, research on the sites was resumed and expanded to include an additional area at the base of the knoll. Abundant chipping debris, unfinished tools, and a large side scraper were found in this new location. On the basis of the exotic raw materials and the stylistic similarities of the side scraper to specimens from the other Paleoindian components, we determined that this newly defined site should be added to the Israel River Complex, and it has been designated as the Jefferson V site. It is the least investigated site in the complex, but has been shown to have two areas of occupation.

The Jefferson IV and V sites have yet to yield sufficient information that would permit a reasonable determination of their function. They served at least as encampments for hunters, and further research may amplify this interpretation or reveal evidence for a broader range of activities. More research is planned for this site in the coming field season.

Return to Jefferson village and turn L. Go W on Rte. 2 about 4.0 mi. Turn L onto Wesson Rd. (just before pipeline pumping station). Go 1.2 mi. and turn L on Mt. Prospect Rd. Continue 1.4 mi. and bear R on Old East Rd. Drive 0.6 mi. and turn R on U.S. Rte. 3. Go N 0.4 mi. and turn R, into Weeks State Park .

STOP 7: WEEKS STATE PARK, Lancaster (R. Boisvert)

This park is also known as the John Wingate Weeks Historic Site. It was the summer retreat for John W. Weeks. Born in Lancaster in 1860 and raised on a nearby farm, Weeks entered the U.S. Naval Academy and eventually prospered as a stockbroker in Massachusetts. He entered local politics there and was appointed US Senator in 1913. Weeks became Secretary of War under Presidents Harding and Coolidge. His principal achievement in government occurred while he was a US Representative with the creation and passage of the Weeks Act. This legislation established the eastern national forest lands system, authorizing the purchase of land to be permanently held and administered for the protection and development of natural resources. The Weeks Act resulted in creation of the White Mountain National Forest and other national forests in the eastern US. Weeks died in 1926 and his children donated the property here on Prospect Mtn. to the state of New Hampshire in 1941.

The lodge and tower at the summit of Prospect Mtn. served as the summer retreat for the Weeks family. The tower was originally a water tower and has been converted to a fire tower. Weeks intentionally constructed it to provide a 360° panorama of the Vermont and New Hampshire countryside. He built the auto road so that local residents and visitors could enjoy the spectacular view. From this vantage point, we can see the Israel River valley and Presidential Range to the southeast. It is a good

location from which to imagine the glaciation of the Israel Valley and damming of glacial Lake Israel during ice retreat. We can also imagine the early postglacial tundra environment that would have attracted large game and Paleoindian hunters to this corridor through the northern White Mountains.

END OF DAY 1

Leave parking lot at Eastgate Motor Inn, exiting onto Cottage St. Drive downhill (toward center of town) for 0.5 mi. and turn R onto Grove St. Go 0.3 mi. and turn L onto Highland Ave. Continue 0.3 mi. and turn L onto Beacon St. Go down steep hill for less than 0.1 mi., and turn L into Marvin's "U-Store-It". Drive around self-storage facility, keeping to L, and park next to excavated bank at far end.

STOP 8: MARVIN PIT, Littleton (W. Thompson)

The following description is abbreviated from Thompson *et al.* (1999). The Marvin Pit is located in a terrace on the south side of the Ammonoosuc River (Figure 6). A 5 m section of silty-sandy till is exposed on the uphill (SE) side of the pit. This till is dark gray (5Y-4/1) and nonoxidized. It is generally nonstratified but contains a few sand lenses. The till is moderately compact with weak fissility. Pebble and cobble-size clasts are abundant, including numerous faceted and striated stones (29% of stone count). There are scattered boulders, many of which are white granite or granodiorite. The till matrix is carbonate-bearing, and carbonate coatings are very common on clast surfaces.

The bedrock along the river bank just southwest of here belongs to the metasedimentary member of the Ordovician Ammonoosuc Volcanics (Lyons *et al.*, 1997). Glacial striations on the ledges indicate ice-flow directions of 125-173°, with the more easterly trends being older. Till-fabric data from the Marvin Pit show a strong preferred orientation in the 320-330° range (Figure 7). This orientation suggests ice flow to the southeast, in agreement with the regional flow trend during the late Wisconsinan glacial maximum.

Leave storage facility, turning L onto Beacon St. Drive 0.3 mi. (cross Ammonoosuc River) to traffic light, and turn R onto Rte.116. Continue 4.1 mi. and turn L onto Douglas Dr. (private rd. leading to Chick's Sand and Gravel). Drive up main gravel rd. for about 0.9 mi. and turn L into pit on downhill side of rd.

STOP 9: INGERSON PIT, Dalton (W. Thompson)

This pit exposes an unusual deposit of thinly-bedded pebble gravel with sand lenses and scattered cobbles and boulders. The following observations are based on a brief visit and are tentative. The gravel includes a great abundance of pink feldspar, suggesting derivation from rotten granite. Identical material occurs in several nearby pits on the other side of the road. The upper part of the deposit (especially the pebble gravel) is quite compact and perhaps has been glacially overridden. Boulders occur on the ground surface in this area.

The local bedrock is mapped as granite of the Alder Brook Pluton (Lyons *et al.*, 1997), though outcrops are rare. This is the same granite body from which glacial ice plucked the countless large boulders scattered across the Ammonoosuc Valley south of here, including the Prospect Boulder Field and Wedick Pit (Stop 13) described below. We suggest that the Alder Brook granite was preglacially weathered or otherwise disintegrated, and that great quantities of grüß were incorporated into the glacial sediments at this locality and elsewhere in the Alder Brook basin. *Questions:* What was the

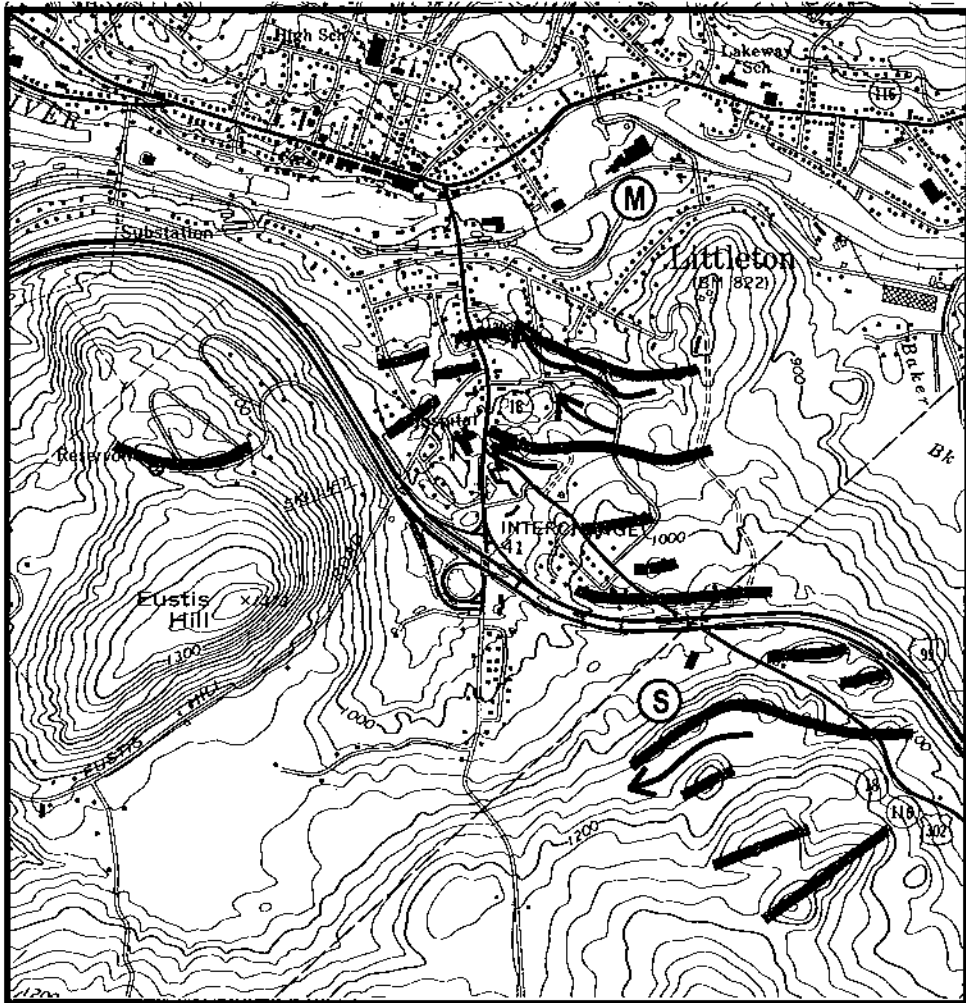


Figure 6. Map of Littleton area, showing moraine ridges forming western part of Bethlehem Moraine complex (thick gray lines) and meltwater channels (arrows). M: Marvin Pit (Stop 8). S: Sleeping Astronomer Moraine (Stop 10). Modified from Thompson *et al.* (1999).

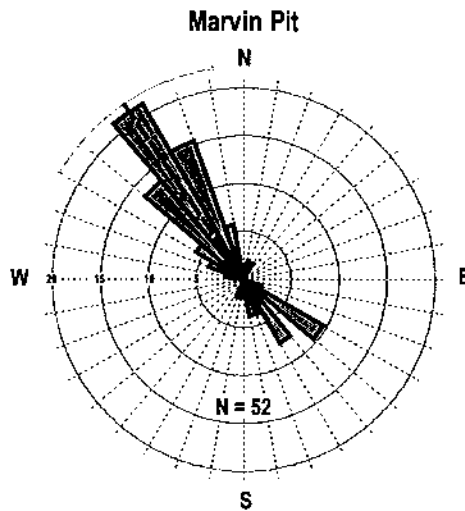


Figure 7. Rose diagram showing fabric of lodgement till in the Marvin Pit, Littleton. From Thompson *et al.* (1999).

origin of this deposit? Is it simply a proglacial meltwater deposit, and was it overridden and overconsolidated by glacial ice? Or is it a strange basal meltout till??

To the north of here, we can see the NE-trending chain of hills called the Dalton Range, which reach elevations of 500-650 m. During the Littleton-Bethlehem Readvance, the profile of the glacier margin must have been steep enough for the late Wisconsinan ice sheet to override these hills and deposit the moraines in the Ammonoosuc Valley. North of Bethlehem village, the ice-surface gradient would have been at least 50 m/km between the crest of the Dalton Range and the youngest moraines just south of the Ammonoosuc River (Thompson *et al.*, 1999).

Looking north from Stop 9, we can see another unusual deposit high on the south face of the Dalton Range. It is a large gravelly outwash fan, which is being worked by Chick's Sand and Gravel. The fan overlies unweathered ablation till that appears identical to the surface till throughout the region. This fan gravel presumably was deposited during the initial retreat of the last ice sheet. It is very compact (overconsolidated) and overlain by scattered large boulders, suggesting it may have been overridden by ice during the Littleton-Bethlehem Readvance. Final recession of the glacier margin to the northwest side of the range was accompanied by a great discharge of meltwater through a gap on Dalton Mountain, carving meltwater channels up to 30 m deep that incise the earlier fan deposit. If this reconstruction is correct, the ice margin may have first receded to the Dalton Range and then readvanced as far as the 12 km between here and the southern limit of the Bethlehem Moraine complex, prior to its final oscillatory retreat (Thompson *et al.*, 1999). Alternatively, deposition of the fan and subsequent overriding of the ice sheet could have occurred late in the history of the moraine complex, requiring a lesser readvance distance.

Return to Rte. 116 and go back to Eastgate Motor Inn in Littleton. From the Eastgate, go E for 0.5 mi. on U.S. Rte. 302. Turn R into Elks Club parking lot, keeping to R, and park out behind building.

STOP 10: SLEEPING ASTRONOMER MORaine, Bethlehem (W. Thompson)

Some of the most clearly defined segments of the Bethlehem Moraine complex can be seen in the town of Littleton and just southeast of town in the adjacent part of Bethlehem. There are numerous moraine ridges in this area, which trend east to northeast. They are up to 36 m high and 0.5 km or more in length. The moraines consist chiefly of till, and large boulders commonly are strewn across their surfaces. Meltwater channels locally occur between adjacent moraines, where water drained along the ice margin.

The moraine at Stop 10 is located just south of U. S. Route 302 (Figure 6). It is over 30 m high on the proximal side and contains many large granitic boulders. One of the larger boulders near the moraine crest at this stop used to be a tourist attraction. It is called "The Sleeping Astronomer" because, when viewed from the downhill side, it resembles the head of a person lying on the ground and gazing skyward (McGoldrick, 1984). A swampy meltwater channel lies on the distal side of the moraine.

Exit parking lot, turn R, and continue E on Rte. 302. Go about 2.0 , and then note that rd. follows crest of low bouldery moraine for next 0.2 mi. (along N side of pond). Continue another 0.4 mi., and Rte. 302 crosses a prominent channel that drained meltwater southward from the glacier margin. Proceed 4.2 mi. E (passing through Bethlehem village) and turn R onto Trudeau Rd. Drive 0.25 mi. and turn R onto private access rd. for gravel pit. Go just 0.05 mi., bear R, and continue into pit.

STOP 11: STONEY'S PIT, Bethlehem (C. Hildreth)

Materials exposed in this pit include: (1) at the base, subaqueous fan deposits of the Beaver Brook Stage of glacial Lake Gale; (2) high on the west wall, lake-bottom sediments deposited in Bethlehem Stage 1 of glacial Lake Ammonoosuc; and (3) overlying all lower natural surfaces, stream-terrace deposits of the late-glacial to postglacial Ammonoosuc River.

This pit was actively excavated in 2000 for a major road project in the area. At that time the lower pit exposed 20+ feet (6 m) of variably bedded, poorly sorted coarse-grained gravel, with some finer gravel and sand, containing cross beds that indicated northward flow. These in turn were overlain by 6 feet (2 m) of horizontally bedded sand and gravel interpreted to be stream-terrace deposits.

The west wall of the pit, at higher elevation, contained a couple of exposures of till overlain directly by 3-9 feet (1-3 m) of well-laminated, horizontal to slightly contorted, thin-bedded silt-clay (varves?), overlain in turn by 10+ feet of planar-bedded fine to medium-grained sand. In part, bedding in the top 6 feet (2 m) of this unit dips north (and thus is interpreted as foresets of Bethlehem Stage 1 of glacial Lake Ammonoosuc (elevation 387 m [1270 ft.]). Recent exposures in this face reveal similar silt-clay beds above till, but at the contact, some of the silt-clay beds are ripped-up fragmented clasts.

Discussion: Exposures of lake-bottom deposits of glacial Lake Ammonoosuc are very rare. The exposures in this pit are the most extensive encountered during mapping in the area (Hildreth, 1984, 2000, 2001). Other exposures include several temporary ones in a couple of small pits along Route 302 near the campground south of Beech Hill, where 6-12 feet (2-4 m) of trough cross-bedded sand and fine gravel overly silt and clay. Silt-clay deposits were also observed in a drainage ditch excavation along Route 3, just east of Haystack Brook and not far from the Gale River 2 spillway of Lake Ammonoosuc. Another small section of silt-clay laminae is exposed at the head of a small landslide scar in material mapped as Stage 5 deposits, high on the bank above the abandoned railroad grade in the gorge east of Pierce Bridge. Lake-bottom deposits exist at depth, as indicated by well-hole data from Flanagan (1996). One well (CFW-53) had 70 feet of sand and gravel over 50 feet of clay over 36 feet of till over bedrock. This well is located south of Route 3 and east of Little River.

Exit pit, turning L on Trudeau Rd., and return to Rte. 302. Turn L, go 2.7 mi., and turn R onto Maple St. in Bethlehem village. Drive 0.75 mi. and turn L onto Bethlehem Water District rd. Continue about 0.5 mi. to where meltwater channel crosses rd., and park nearby.

STOP 12: LAKE AMMONOOSUC SPILLWAY CHANNELS, Bethlehem (C. Hildreth)

The outlet for the Bethlehem 2 Stage of glacial Lake Ammonoosuc, at 369 m (1211 ft) altitude, consists of several wide WNW-trending swampy channels carved in till and lined with abundant surface boulders in places. These channels coalesce into a single broad swamp just west of the Bethlehem sewer beds, which are partly constructed on the proximal (eastern) end of at least one of the channels. The road to the facility crosses one of the channels just before reaching the building. The broader channel can be viewed by following a gravel path westward around the gated area for several hundred feet.

Return to Maple St. Turn R and go 0.75 back to Rte. 302. Turn R onto Rte. 302 and drive 0.35 mi. Turn R onto Prospect St. and go 1.1 mi. to **Bretzfelder Park** on L. The picnic shelters in the park overlook Barrett Brook, which here follows one of the spillway channels for glacial Lake Ammonoosuc.

From Bretzfelder Park, go N on Prospect St. for 0.45 mi. and bear R at fork. For the next 0.4 mi., Prospect St. follows the crest of one of the **Midacre Farm Moraine**, which is one of the larger segments of the Bethlehem Moraine complex (Stop 4 on NEIGC trip by Thompson *et al.* [1996]). The ridge trends northeast, reaches an elevation of about 399 m, and is at least 24 m high at this point. When this moraine segment was deposited it stood well above the adjacent Bethlehem 2 Stage of Lake Ammonoosuc, which was controlled by the nearby B2 spillway at 369 m. The ridge appears to consist mainly of till with scattered boulders, although a small water-laid deposit was mapped at the west end (Flanagan, 1996). Till is exposed in a cut bank along Prospect Street just northeast of here, and in an abandoned pit next to nearby Willson Road. The till in this area is sandy, stony, and olive-gray, which is typical of the fresh late Wisconsinan surface till in the White Mountains. A well located just northeast of here on Prospect Street penetrated 35 m of surficial sediments before reaching bedrock.

Continuing NE just past Willson Rd., Prospect St. descends across the east side of a high ridge that Goldthwait (1916) and Crosby (1934) mapped as part of the Bethlehem Moraine complex. The 417-m summit of this hill stands about 77 m above the contact with a sandy terrace downhill from here. When the leaves drop from the trees in the fall, the field of large boulders covering the hillside is an impressive sight. This is the **Prospect Boulder Field** visited during the 1996 NEIGC trip. Nearly all of these boulders are white-weathering granite. They were glacially plucked from a rock body that underlies this area, called the "Hedgehog Hill granite" by Billings (1956) and now known as the "Alderbrook pluton" (E. Boudette, pers. comm., 1996). Because of the extreme abundance of the granite boulders and the high steep slope, it was expected that bedrock outcrop would be found on this hillside. No ledge was seen, however, and a well drilled on the crest of the ridge encountered 41 m of surficial sediments overlying bedrock (N.H. Water Well Board, record no. 025.0127, Willson Heights Rd.). The driller's log reported 24 m of "gravel" [loose till?] overlying 17 m of till, followed by "salt and pepper rock" [granite]. This ridge is much higher than other segments of the Bethlehem Moraine complex, and does not have the usual east-northeasterly trend. The oval shape of the contours and smooth proximal till slope resemble many other hills in the area, where glacial processes have accreted thick till deposits over bedrock knobs. Thompson *et al.* (1996) proposed that the ridge may be a composite of earlier lodgement till (recorded in the lower part of the driller's log) that is draped by the bouldery till deposits typical of the Bethlehem Moraine. A lower bouldery ridge juts east from the large hill. This ridge may be a crosscutting moraine built by a small ice lobe in the Wing Road area.

Continue on Prospect St., cross bridge on Ammonoosuc River, and turn R onto Wing Rd. Bear R at fork. Follow river for 0.45 mi. and turn L onto private driveway for Wedick residence. Continue 0.35 mi., past house in woods and then downhill into gravel pit.

STOP 13: WEDICK PIT, Bethlehem (W. Thompson)

The Wedick Pit (see location on Figure 3) extends east-southeast from Wing Road. It has been dug into a terrace with an original surface elevation of about 324 m (very slightly lower than the Wing Road Stage of Lake Ammonoosuc). When the Wedick Pit was visited during the 1996 NEIGC conference, a distinctive feature was the presence of many large boulders which were embedded in and rested upon the surface of a till deposit. These boulders were commonly 2-3 m across, the majority being white to pink granite and gneissic granite. Most of them were recently removed and used for crushed stone or road fill, although a few large examples still remain. The depth to bedrock at two nearby sites (north and south of the pit) is 40-41 m, as indicated by well logs. However, a seismic profile and observations by the owner indicate that the bedrock surface is at a much shallower depth beneath the floor of the Wedick Pit.

The lowest stratigraphic unit presently seen in the Wedick Pit is till, which is exposed in the long water-filled trench in the pit floor. According to the owner, a clayey and very dense till, as well as lacustrine clay, was excavated from the deepest part of this trench, which has since been backfilled. The till on the north side of the trench shows much deformation and contains numerous silt-sand lenses and sand clasts. In the end of the pit close to Wing Road, a small section exposed laminated diamict consisting of alternating sandy and muddy layers with scattered pebbles and cobbles. In 1995, sections just south of the trench showed the following composite stratigraphy:

2.6 m fluvial pebbly sand and pebble-cobble gravel with cross-bedding dipping NNW

----- sharp contact -----

1.4 m oxidized, stony, olive-colored diamict (5Y-5/3); mottled appearance resulting from many lenses of bluish-gray silt-sand and coarser white sand, which are greatly deformed due to flowage (?); contains a few lenses of pebble gravel

0.8 m diamict; mottled but mostly gray (5Y-5/1); less stony than overlying unit; contains subhorizontal lenses of white sand (not greatly deformed) and muddy very fine sand

Judging from these exposures, much of the till at this locality was deposited in the presence of abundant meltwater, and probably into a glacial lake. We infer that this water-laid till accumulated where the late Wisconsinan ice margin was in contact with Lake Ammonoosuc. In the eastern part of the pit, there are exposures of lacustrine sand with foreset beds and current ripples indicating transport generally to the north. This deltaic sediment was deposited by the early Ammonoosuc River flowing into one of the final stages of Lake Ammonoosuc.

The youngest unit in the Wedick Pit is the fluvial sand and gravel (which locally has been removed). This unit is exposed south of the trench, where it directly overlies till in the sequence described above. It is believed to be a late-glacial or early postglacial terrace deposit of the Ammonoosuc River, formed as the river began to rework glacial sediments in the valley and cut down to its present level.

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**FRIENDS OF THE PLEISTOCENE REUNIONS
1934-2002**

Reunion	Leaders	Area
1. 1934	George White / J.W. Goldthwait	Durham to Hanover, NH
2. 1935	Dick Flint	New Haven to Hartford, CT
3. 1936	Kirk Bryan	SE Rhode Island to Cape Cod, MA
4. 1937	J.W. & Dick Goldthwait / Dick Lougee	Hanover to Jefferson, NH
5. 1938	Charlie Denny / Hugh Raup	Black Rock Forest, NY
6. 1939	Paul MacClintock / Meredith Johnson	Northern NJ (drifts)
7. 1940	Kirtley Mather / Dick Goldthwait	Western Cape Cod, MA
8. 1941	John Rich	Catskill Mtns., NY
<i>1942-45</i>	<i>no meetings during war years</i>	
9. 1946	Lou Currier / Kirk Bryan	Lowell-Westford area, MA
10. 1947	Earl Apfel	Eastern Finger Lakes, NY
11. 1948	D.F. Putnam / Archie Watt / Roy Deane	Toronto to Georgian Bay, ONT
12. 1949	Paul MacClintock / John Lucke	NJ ("Pensauken problem")
13. 1950	O.D. Von Engeln	Central Finger Lakes, NY
14. 1951	John Hack / Paul MacClintock	Chesapeake, MD (soils/stratigraphy)
15. 1952	Dick Goldthwait	Central OH (tills)
16. 1953	Lou Currier / Joe Hartshorn	Ayer quad, MA (outwash sequences)
17. 1954	Charlie Denny / Walter Lyford	Wellsboro-Elmira-Towanda, PA-NY
18. 1955	Paul MacClintock	Champlain lake and sea, NY
19. 1956	Nelson Gadd	St. Lawrence Lowland, QUE
20. 1957	Paul MacClintock / John Harris	St. Lawrence Seaway, NY
21. 1958	John Hack / John Goodlett	Appalachians, Shenandoah, VA

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| 22. 1959 | Alexis Dreimanis / Bob Packer | Lake Erie, ONT (till bluffs) |
| 23. 1960 | Ernie Muller | Cattaraugus Co., western NY |
| 24. 1961 | Art Bloom | SW Maine (marine clay; ice margins) |
| 25. 1962 | Cliff Kaye / Phil Schafer | Rhode Island (Charleston Moraine etc.) |
| 26. 1963 | Hulbert Lee | Lower St. Lawrence Lowland, QUE |
| 27. 1964 | Cliff Kaye | Martha's Vineyard, MA |
| 28. 1965 | Joe Upson | Northern Long Island, NY |
| 29. 1966 | Nick Coch / Bob Oaks | Southeast VA (scarps; stratigraphy) |
| 30. 1967 | Hal Borns | Eastern ME (moraines; glaciomarine) |
| 31. 1968 | Carl Koteff / Bob Oldale / Joe Hartshorn | Eastern Cape Cod, MA |
| 32. 1969 | Nelson Gadd / Barrie McDonald | Sherbrooke area, QUE |
| 33. 1970 | Dick Goldthwait / George Bailey | Mt. Washington area, NH |
| 34. 1971 | Gordon Connally | Upper Hudson Valley, Albany, NY |
| 35. 1972 | Art Bloom / Jock McAndrews | Central Finger Lakes, NY |
| 36. 1973 | Don Coates / Cuchlaine King | Susquehanna-Oswego Valleys, NY-PA |
| 37. 1974 | Bill Dean / Peter Duckworth | Oak Ridges-Crawford Lake, ONT |
| 38. 1975 | George Crowl / Gordon Connally /
Bill Sevon / Les Sirkin | Lower Delaware Valley, PA |
| 39. 1976 | Bob Jordan / John Talley | Coastal Plain, DE |
| 40. 1977 | Bob Newton | Ossipee area, NH |
| 41. 1978 | Denis Marchand / Ed Ciolkosz /
Milena Bucek / George Crowl | Central Susquehanna Valley, NY |
| 42. 1979 | Jesse Craft | NE Adirondack Mtns., NY |
| 43. 1980 | Bob LaFleur / Parker Calkin | Upper Cattaraugus, Hamburg, NY |
| 44. 1981 | Carl Koteff / Byron Stone | Nashua Valley, MA |
| 45. 1982 | Pierre LaSalle / Peter David /
Michelle Bouchard | Drummondville, QUE |

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| 46. 1983 | Woody Thompson / Geoff Smith | Augusta-Waldoboro area, ME |
| 47. 1984 | Peter Clark / J.S. Street | St. Lawrence Lowland, NY |
| 48. 1985 | Ed Evenson / Jim Cotter / Dave Harper / Carl Koteff / Jack Ridge / Scott Stanford / Ron Witte | Great Valley, NJ-PA |
| 49. 1986 | Tom Lowell / Steve Kite | Northernmost ME |
| 50. 1987 | Carl Koteff / Janet Stone / Fred Larsen / Joe Hartshorn | Connecticut Valley-Lake Hitchcock, CT-MA |
| 51. 1988 | Ernie Muller / Duane Braun / Bill Brennan/ Dick Young | Genesee Valley, NY |
| 52. 1989 | Pierre LaSalle / Andree Blais / Denis Demers / Michel Lamothe / Bill Shilts | Mid St. Lawrence Lowland, QUE |
| 53. 1990 | Ralph Stea / Bob Mott | Halifax region, NS |
| 54. 1991 | Jack Ridge | Western Mohawk Valley, NY |
| 55. 1992 | Bob Dineen / Eric Hanson / Bob LaFleur / Dave Desimone | Lower Mohawk Valley, NY |
| 56. 1993 | Carol Hildreth / Richard Moore | Contoocook-Souhegan-Piscataquog Valleys, NH |
| 57. 1994 | Duane Braun / Ed Ciolkosz / Jon Inners / Jack Epstein | Eastern PA |
| 58. 1995 | Woody Thompson / Tom Davis / John Gosse / Bob Johnston / Bob Newton | Portland-Sebago Lake-Ossipee Valley, ME |
| 59. 1996 | Hal Borns / Chris Dorion / Joe Kelley / Karl Kreutz / Dave Smith / Woody Thompson/ Rick Will | Glaciomarine deposits, eastern ME |
| 60. 1997 | Scott Stanford / Ron Witte | Northern New Jersey |
| 61. 1998 | Les Sirkin | Long Island, NY |
| 62. 1999 | Ben Marsh | Periglacial landscapes, central PA |
| 63. 2000 | Julie Brigham-Grette / Tammy Rittenour / Janet Stone / Jack Ridge / Al Werner / Dena Dincauze / Ed Klekowski / Richard Little | Glacial Lake Hitchcock, MA |

64. 2001 Najat Bhiry / Jean-Claude Dionne / Martine Clet / Serge Occhietti / Jehan Rondot Quebec City region, QC
65. 2002 Woody Thompson / Carol Hildreth / Dick Boisvert / Chris Dorion / Brian Fowler Northern White Mtns., NH