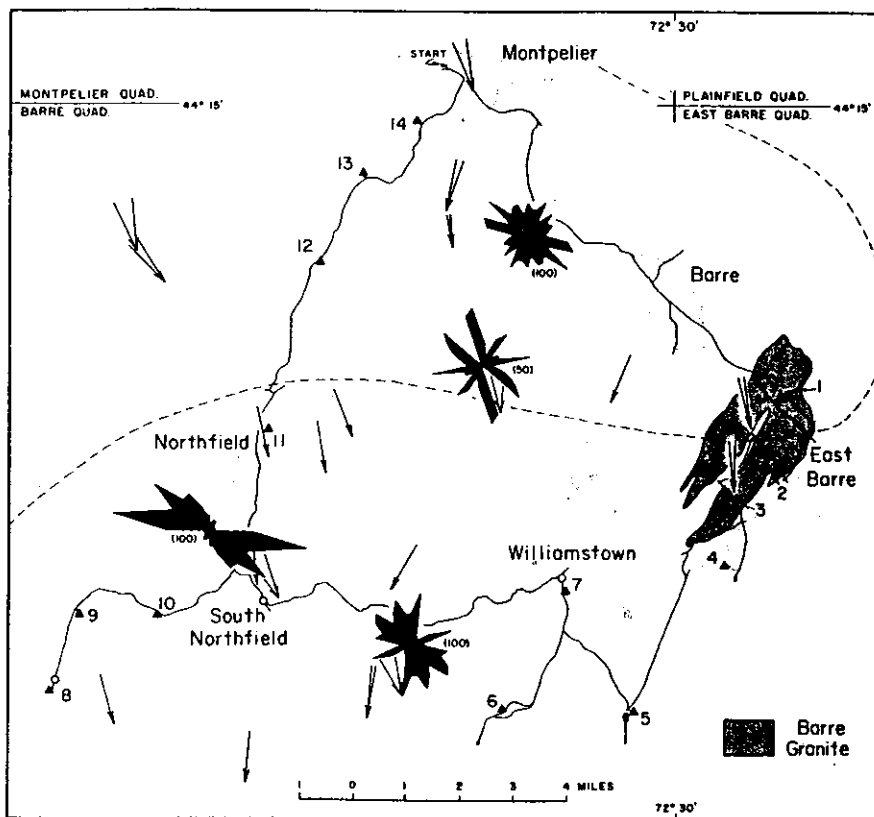


Glacial, Late-Glacial and Postglacial History of Central Vermont

Guidebook for the 66th Annual Meeting of the Northeast
Friends of the Pleistocene

Montpelier, Vermont
May 23-25, 2003

Field Trip Leaders: Frederick D. Larsen
Stephen F. Wright
George E. Springston
Richard K. Dunn



DEDICATION

Ernest H. Muller

The Friends of the Pleistocene, an organization (or non-organization) without officers and dues, is greatly appreciative of Ernie Muller's service as our guiding mentor. Ernie has been the one to recruit, cajole, and twist the arms of trip leaders for over a decade. It has been hard to say 'NO'. Friends trips over the last decade reflect Ernie's grasp of current research and field work in the Northeast and his undying interest in the work of others from western New York to Nova Scotia.

Ernie has been an excellent listener and supporter of his students and colleagues. He embraces new ideas with a respectful professional skepticism. As his former students will certainly attest, his criticisms are challenging but always constructive and his thoughtful analyses and gentle suggestions have given us direction and opened our minds to a multitude of new ideas. As the sign in his office at Syracuse reads, "I said maybe and that's final." For so many of us, Ernie epitomizes the ideal of gentleman-scholar and we are proud to call him a colleague and Friend.

Thank you Ernie!



Cover: Directional features in central Vermont: striations, till fabrics and a crag-and tail (at 5) (Larsen, 1972). Dashed line is edge of "Burlington drift" of Stewart and MacClintock (1969, 1970).

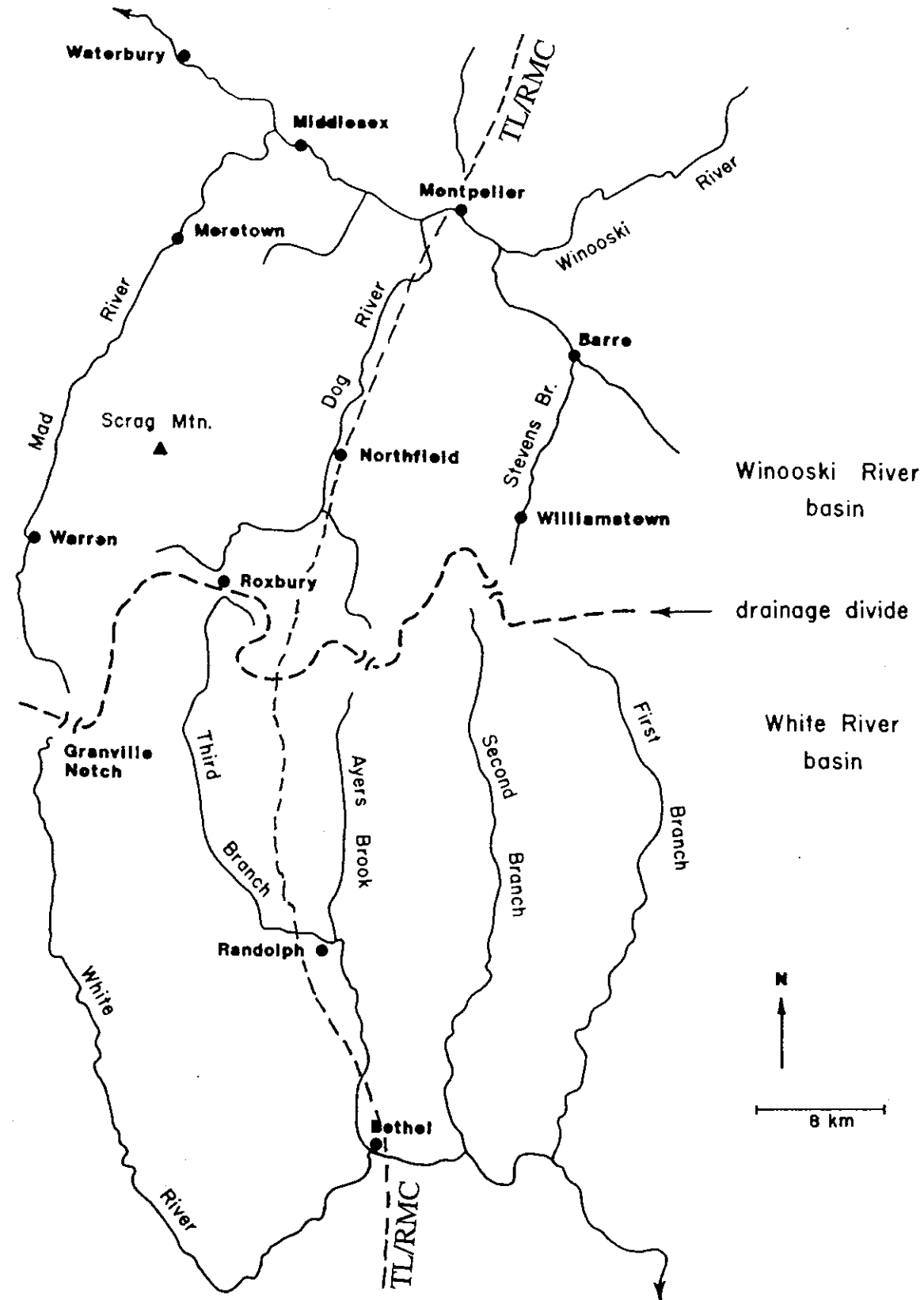


Figure 1. Stream drainage pattern in central Vermont. Heavy dashed line is the drainage divide between northwest-flowing Winooski River and southeast-flowing White River. Light dashed line is Taconian Line/Richardson Memorial Contact (TL/RMC).

GLACIAL, LATE-GLACIAL AND POSTGLACIAL HISTORY OF CENTRAL VERMONT

by

Frederick D. Larsen, Norwich University (retired)
 Stephen F. Wright, University of Vermont
 George E. Springston, Norwich University
 Richard K. Dunn, Norwich University

INTRODUCTION

Welcome to the 66th annual meeting of the Northeast Friends of the Pleistocene. We are pleased to host the first Friends meeting ever held in Vermont. There have been several recent glacial geology field trips in central Vermont associated with NEIGC (New England Intercollegiate Geological Conference). NEIGC field trips were led by Fred Larsen (1999a,b) in the Montpelier quadrangle, Stephen Wright (1999a,b) in the Barre West quadrangle and George Springston and George Haselton (1999a,b) in the St. Johnsbury quadrangle. Each of these field trips resulted from detailed mapping of surficial deposits in 7.5-minute quadrangles that was supported by the U. S. Geological Survey and the Vermont Geological Survey. Also in 1999, Bierman, Wright and Nichols led a regional trip from Burlington to Bolton, Smugglers Notch and Jeffersonville. Earlier NEIGC trips related to central Vermont were led by Larsen (1972, 1987a, 1987b) and by Ackerly and Larsen (1987). Since 1999, surficial mapping by Larsen and others (2002) has been completed in the drainage basin of the Third Branch of the White River (includes the Randolph quadrangle). Mapping in the drainage basin of the Mad River by George Springston, Richard Dunn and Nathan Donahue currently is in progress.

Of 14 possible stops described in this guidebook (not all will be made) nine are new and only five have been visited during recent NEIGC field trips (Plate 1, separate Field Trip Map). The area traversed on this field trip on the first day lies on the following Vermont 7.5-minute U. S. Geological topographic maps: Randolph Center (Stop 1), Brookfield (Stops 2 and 3), Barre West (Stops 4, 5 and 6), Barre East (Stop 7), Montpelier (Stops 8 and 9), on the second day Bolton Mountain (Stop 10), Roxbury (Stops 11 and 12), and Randolph (Stops 13 and 14).

An important geologic contact in the bedrock is known variously as the Taconian Line (TL, Hatch, 1982), the Richardson Memorial Contact (RMC, informal designation) or the Dog River Fault Zone (DRFZ, Westerman, 1987). The line trends north-south and passes under the Vermont State Capitol in Montpelier and through or close to the villages of Northfield, Randolph and Bethel (TL/RMC, Fig. 1). The rocks west of the line are Cambrian and Ordovician in age and mainly consist of greenish-gray, chlorite-rich quartzite, phyllite, schist and dark green greenstone. East of the line the age of bedrock is under debate, but is in the range of Ordovician to Devonian, and the character of the rocks is much different. The Waits River Formation is composed of interbedded light to dark gray calcareous quartzite and phyllite and the Northfield Formation is made up mostly of dark to light gray phyllite and slate. The rocks to the west of the Taconian Line were deformed and metamorphosed mainly during the Taconic orogeny but were also affected during the Acadian orogeny. The Waits River and Northfield Formations were metamorphosed and deformed during the Acadian orogeny, and intruded by the Barre Granite, which has a radiometric age of 380 Ma.

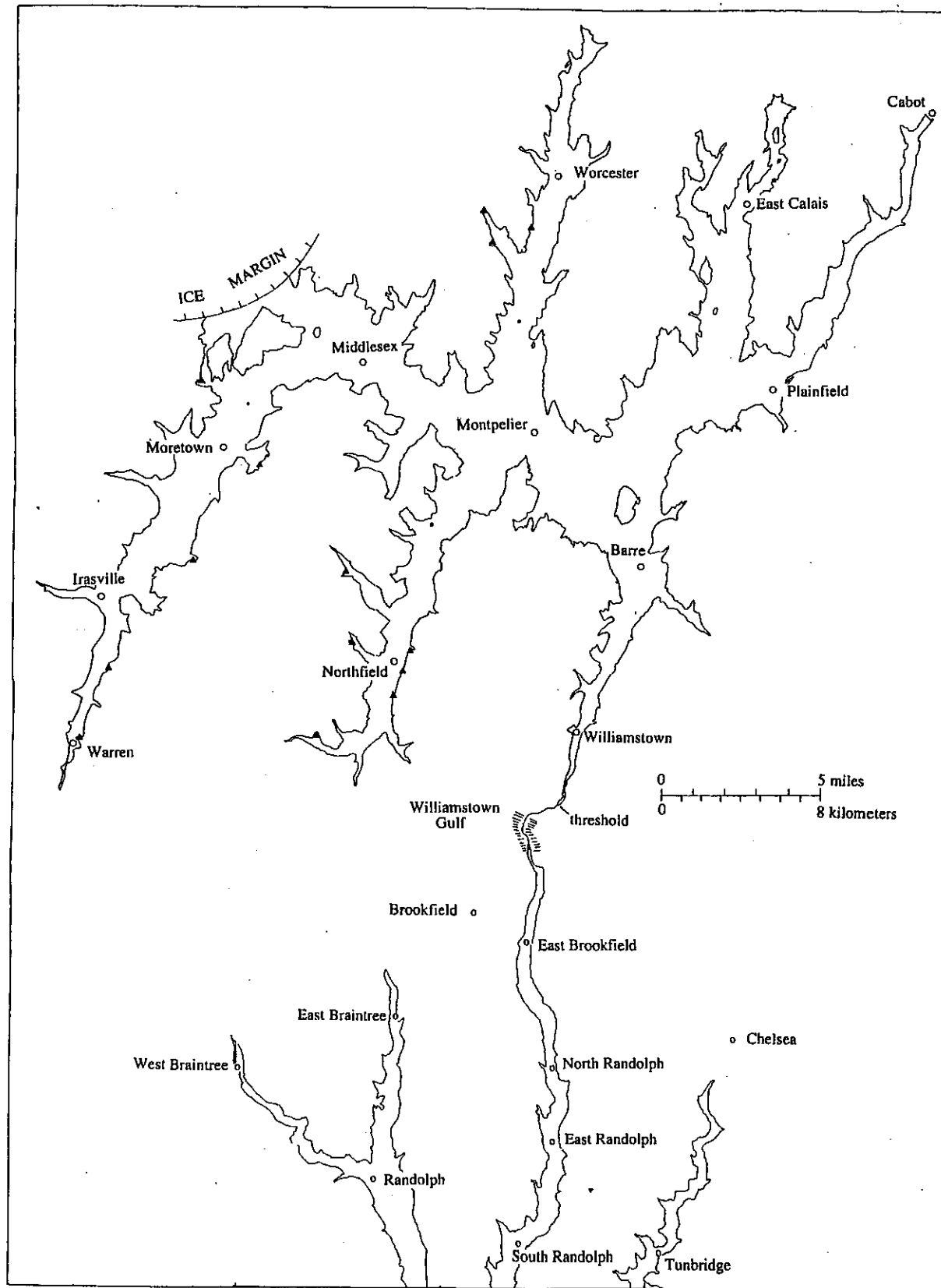


Figure 2. Projected shorelines of Lake Winooski (top) and Lake Hitchcock (bottom) using tilt of 4.74 ft/mi (0.90 m/km) after Koteff and Larsen (1989). Lake Hitchcock filled south-draining valleys of the White River: from right to left the First, Second and Third Branches of the White River. Ayers Brook valley drains from East Braintree to Randolph.

metamorphosed and deformed during the Acadian orogeny, and intruded by the Barre Granite, which has a radiometric age of 380 Ma.

Erosion has produced a crude trellis drainage pattern characterized by alternate linear ridges and subsequent valleys that trend north-northeast/south-southwest (Fig. 1). Local relief is on the order of 1000 ft with Scrag Mountain at 2,911 ft ASL in the Northfield Range and Mt. Hunger at 3,539 ft ASL in the Worcester Range north-northwest of Montpelier being the higher peaks. Further afield, Mt. Mansfield, located 23 mi northwest of Montpelier, at 4,393 ft ASL is the highest peak in Vermont.

Critical to an understanding of the deglacial history of central Vermont is the drainage divide separating north-flowing tributaries of the Winooski River from south-flowing tributaries of the White River (Fig. 1). In the larger view this divide separates the St. Lawrence River and Connecticut River drainage basins. In late-glacial time, low spots on the drainage divide acted as thresholds for proglacial lakes that formed after the ice margin had retreated north of the divide. The important thresholds that controlled local ice-marginal lakes are from west to east: (1) Granville Notch, 1,410 ft ASL, (2) Roxbury, 1,010 ft ASL, and (3) 2.5 mi south of Williamstown, 915 ft ASL. A fourth threshold, located at the head of Brookfield Gulf (Ayers Brook) 7.5 mi south of Northfield, has an elevation of 1,430 ft ASL. A small proglacial lake that developed north of this threshold has no major deposits associated with it, and did not play a major role in the deglacial history (Larsen, 1987b).

During the Wisconsin glacial (90,000 to 12,000 B.P.), New England was depressed by the weight of the ice sheet. When the weight of the ice sheet was removed during deglaciation, the land rebounded to its present position. The amount and direction of uplift was measured by Koteff and Larsen (1989) to be 4.74 ft/mi (0.9 m/km) to N20.5°W-N21°W using topset/foreset contacts in deltaic deposits of glacial Lake Hitchcock in the Connecticut River drainage basin. Lake Hitchcock extended north as far as Williamstown Gulf in the valley of the Second Branch of the White River (Stops 1 and 2), north past East Braintree in Ayers Brook valley (Stop 13), and north to West Braintree in the valley of the Third Branch of the White River (Fig. 2). Assuming that any smaller glacial lake in western New England experienced the same rebound, a projection for a glacial lake in the Winooski River drainage basin was drawn starting at the 915-ft (279-m) threshold south of Williamstown and extended into the Winooski basin (Fig. 2) (Larsen, 1987b, 1999b). The projection falls at the break-in-slope (approximate topset/foreset contact) of more than 15 deltas shown by black triangles in the Winooski basin.

Glacial lakes filled the Winooski River basin during both the advance and retreat of the last (Laurentide) ice sheet. Mapping of surficial deposits in central Vermont has revealed several locations with till overlying lake-bottom sediments (Fig. 3). The senior author wrote in 1999, "Lacking other evidence for a local readvance by active ice, these till-over-varves sections are interpreted to be related to the Late Wisconsin ice advance". During the 1999 NEIGC field trip C1 at Culver Brook, Stop 5B, organic material was collected by Jack Ridge (Tufts University) that has an AMS date of $11,900 \pm 50$ ^{14}C years BP thus documenting the Middlesex readvance (Larsen, 2001). This is the same age assigned by Ridge and others (1999) for the Littleton-Bethlehem readvance at the Comerford Dam site in Barnet, Vermont. How many of the till-over-

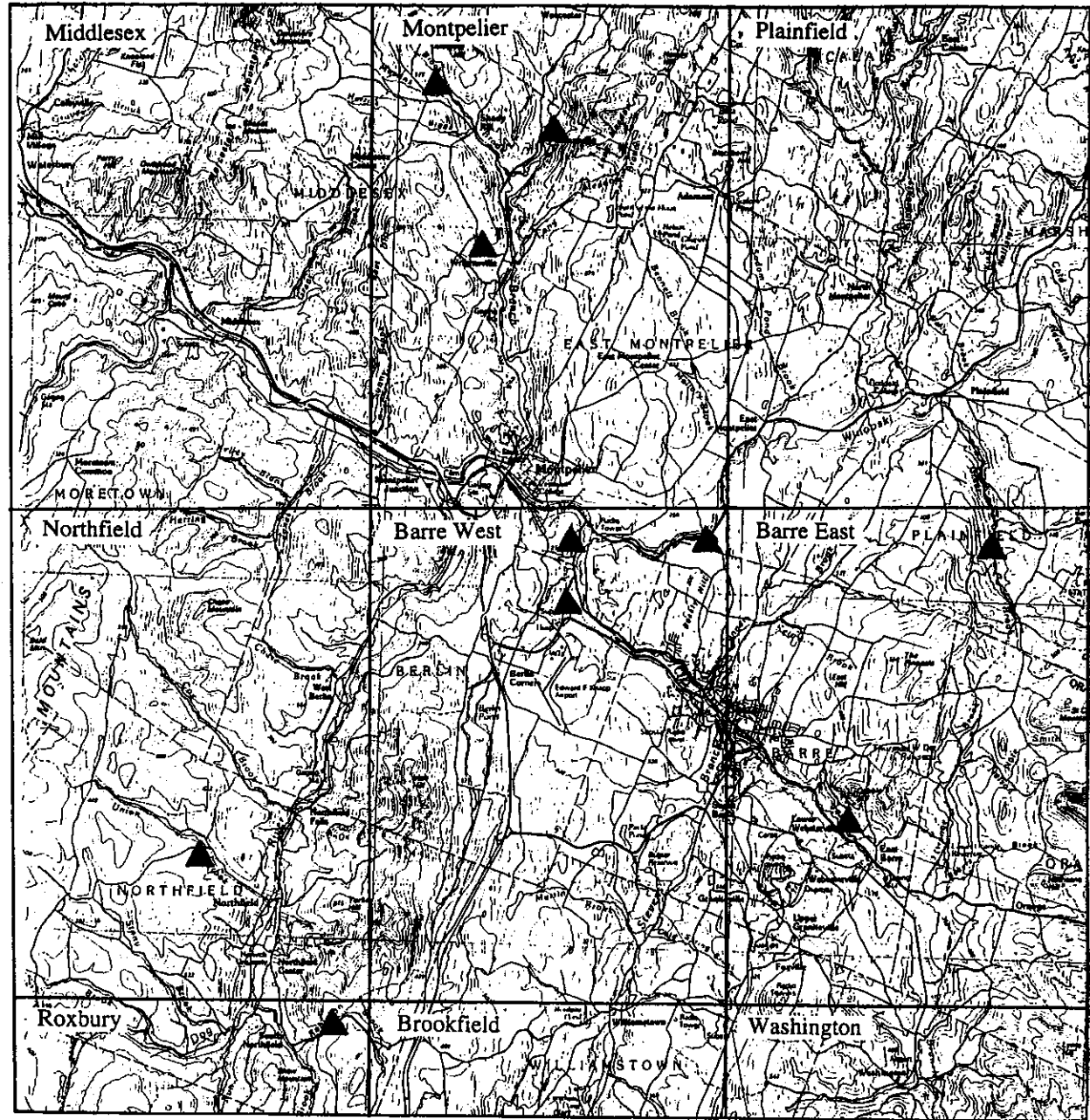


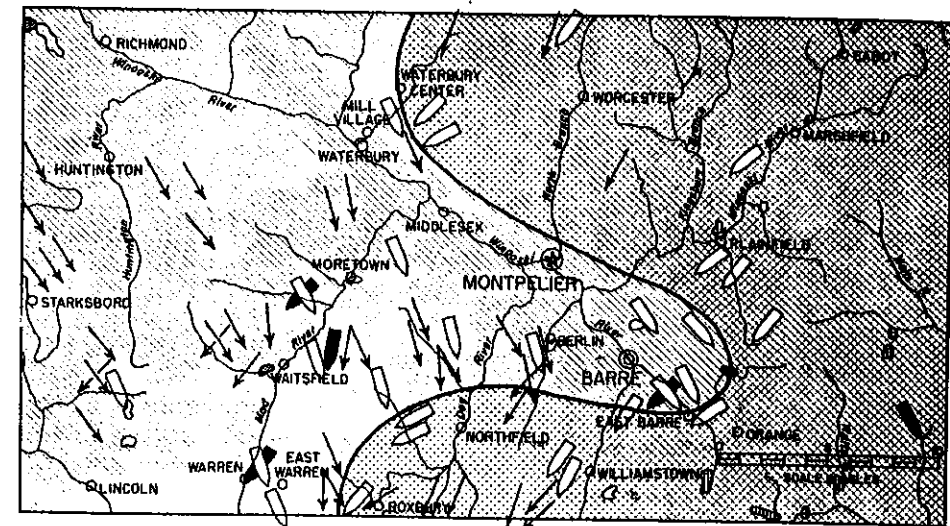
Figure 3. Solid triangles denote sites in central Vermont where till overlies clay-silt varves or stratified sand and gravel. Three sites in the Montpelier quadrangle are related to the Middlesex Readvance at 11,900 BP.

varves sites shown on Figure 3 are related to the Middlesex Readvance or to the main Late-Wisconsinan ice advance remains a question that can be addressed during this Friends trip.

DIRECTION OF ICE MOVEMENT

On their Surficial Geologic Map of Vermont, Stewart and MacClintock (1970) displayed an inset map showing three drift sheets. The oldest drift sheet, the Bennington, was presumably formed by ice moving southeast in southern Vermont. A second drift sheet, the Shelburne, presumably was formed by ice moving southwest throughout much of eastern Vermont. The youngest drift sheet, the Burlington, was formed by ice moving from northwest to southeast in northwest Vermont. The boundary between the Burlington and Shelburne drift sheets passed through the Montpelier-Barre area (Fig. 4). A compilation of striations in New England by James W. Goldthwait (Fig. 5, in Flint, 1957) indicates southwest, south and southeast movement of ice in eastern Vermont. A compilation of indicator fans in New England (Fig. 6, also in Flint, 1957) shows glacial movement between south and southeast. Many 15-minute bedrock maps published by the Vermont Geological Survey have striations plotted on them. Woodland (1965) shows only southeast-trending striations in the Burke quadrangle. Dennis (1956) plotted both southwest- and southeast-trending striations in the Lyndonville quadrangle. The average direction of 21 plotted striations in the Hardwick quadrangle is S 9.5°E (Konig and Dennis, 1964). None of this published data as it relates to the area of the Shelburne drift was used by Stewart and MacClintock in 1970.

This interesting episode regarding the glacial history of Vermont would not be recycled here, except we continue to meet pit owners, well drillers and others who "know" about the "three glaciations" or "three drift sheets". Sometimes the printed work dies hard.



INDEX MAP SHOWING THE BURLINGTON DRIFT BORDER IN THE MONTEPELIER REGION

Figure 4. Map showing the Burlington drift border in central Vermont (Fig. 15, in Stewart and MacClintock, 1969).

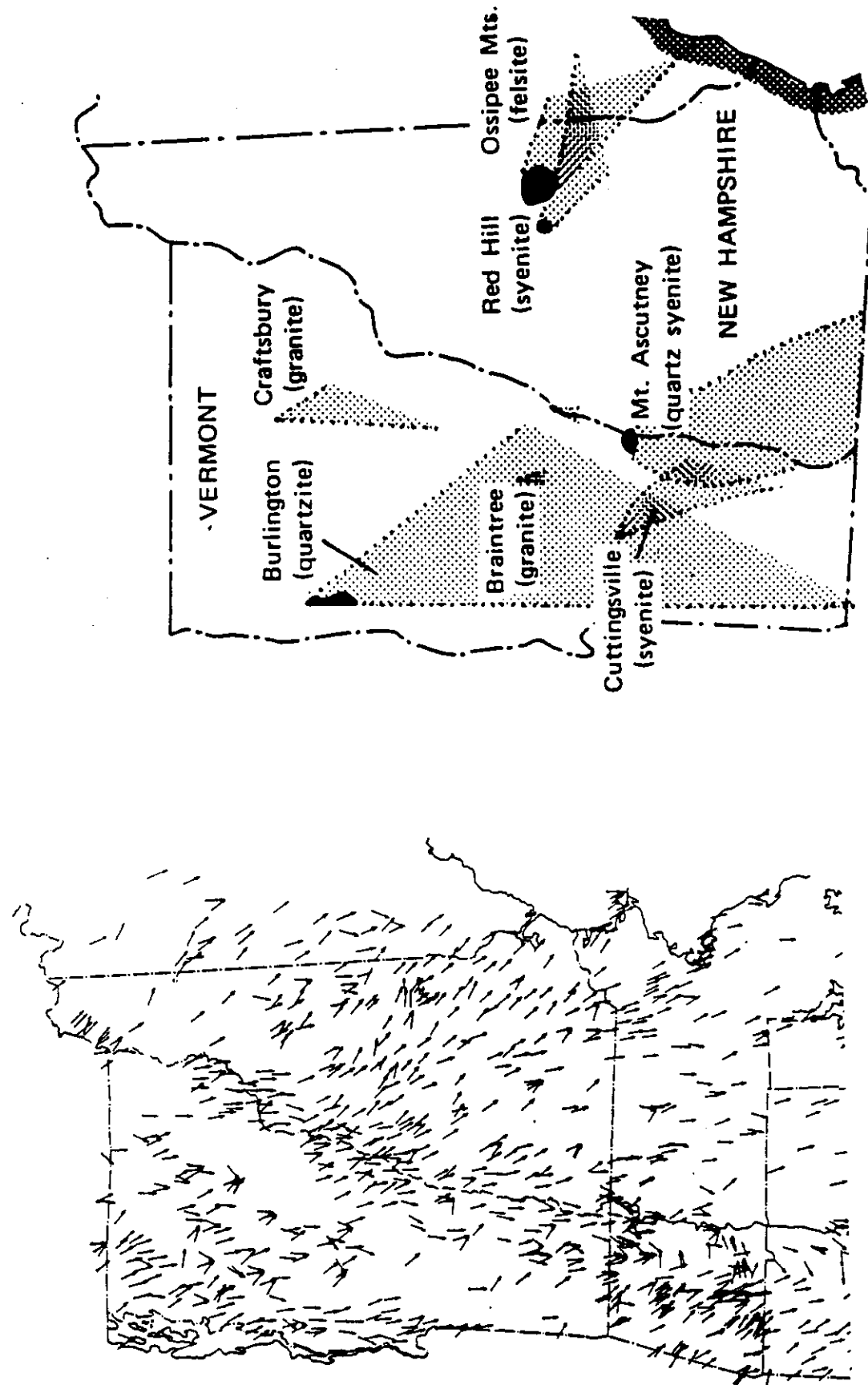


Figure 5. Compilation of striations in Vermont and New Hampshire by James W. Goldthwait (portion of Fig. 5-3 in Flint, 1957, p. 60).

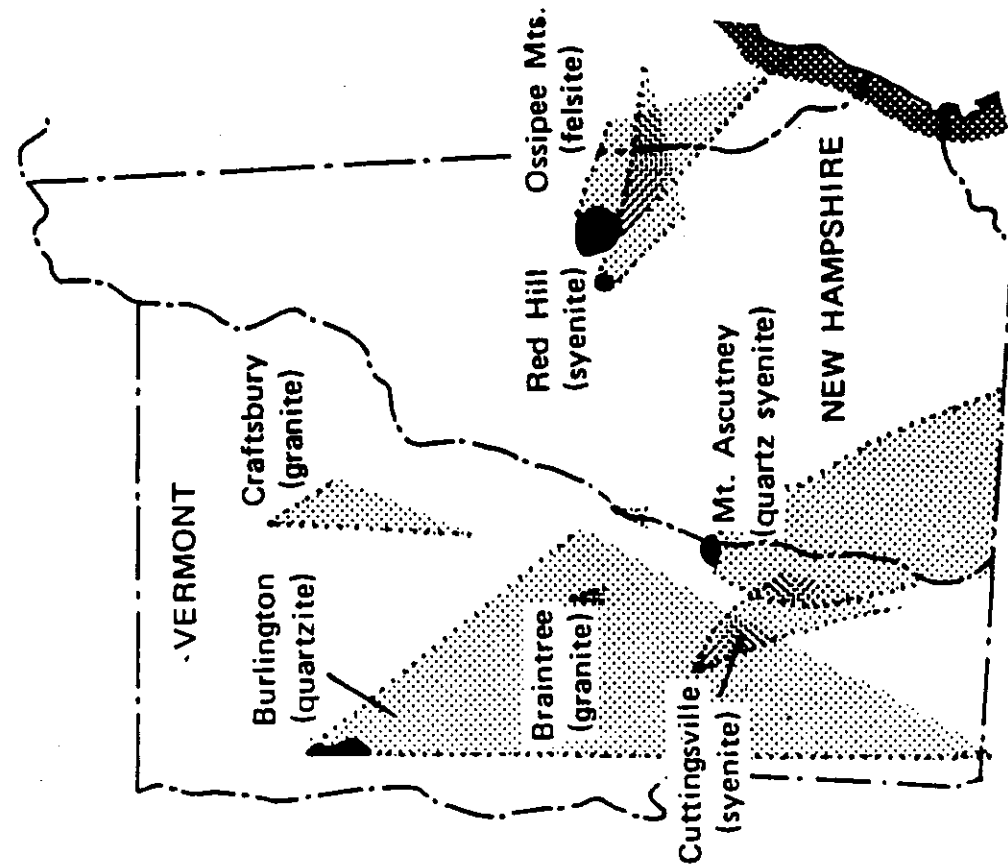


Figure 6. Indicator fans in Vermont and New Hampshire (portion of Fig. 7-19 in Flint, 1971, p. 178).

“How the Great Pebble Campaign was born” (personal account, F. D. Larsen)

In 1967, I attended the GSA Northeast Section meeting in Boston. Dr. David P. Stewart presented a map showing 2 drift sheets in central Vermont (Fig. 1). An older Shelburne drift was formed by ice moving from northeast to southwest and a younger Burlington drift formed by ice moving from northwest to southeast. I was suspicious of the map because (1) the lobe of Burlington drift was not symmetrical over the Winooski valley and (2) having hiked along the crest of the southern Worcester Range (between Waterbury Center and Worcester) I knew that the striations there trend S35 to 40°E in an area of the Shelburne drift. I also knew that there was a simple test of the direction of ice movement in central Vermont because the margin of the Burlington drift at East Barre cut across the middle of the Barre pluton. The distribution of erratics and pebbles of light-colored Barre Granite on terrain of dark-colored metamorphic rocks should point the way.

During the summer of 1967, Eugene “Skip” Rhodes, U. Mass. '68, and I mapped the indicator fan of Barre Granite pebbles. We collected 100 pebbles at each site then plotted the percent of granitic pebbles at each site on the regional map (Fig. 7). The modal direction of ice movement was S17°E from the Barre pluton. The indicator fan map was published in the NEIGC guidebook (Larsen, 1972) and I sent a copy of the map to Dr. Stewart. At a Penrose Conference on the Pleistocene Stratigraphy of the Northeast in the fall of 1974, Dr. Stewart gave a presentation of the three drift sheets in Vermont using the same maps that he had used in 1967. I reminded him that the Barre Granite indicator fan had a trend of S17°E in an area of the Shelburne drift. He said, “I’ll move my line over”. I decided to study indicator fans in eastern Vermont. The Great Pebble Campaign was born.

We did not have a geology major at Norwich University at that time and I decided to use my unbiased physical geology students in a 3-week lab sequence to (1) collect the pebbles, (2) identify the pebbles and (3) contour and interpret the data. Earlier, I had collected a few pebble counts around the Braintree pluton so that area northwest of Randolph, Vermont, became the subject of the first “Great Pebble Campaign”. Over the next decade, physical geology students at Norwich mapped eight more indicator fans in Vermont and New Hampshire (Fig. 8). All displayed transport directions between due south and southeast in the area of the Shelburne drift.

The students had a great time getting off campus. Norwich being a military college, freshmen normally were not allowed to leave campus until Thanksgiving break. Enough upperclassmen had cars so that three freshmen would pile in with the upperclassman and go off with shovels, cardboard boxes, and a topographic map with collection sites clearly marked. Often, there were beer bottles in with the pebbles. We did have a coverup that we called “pebblegate”. Four young men came back with samples containing between 12 and 18 per cent granitic clasts. I told them that I was doubtful that they had followed instructions and for one week they stonewalled me. Finally, I sat them down and asked them to level with me. They said they had collected all the pebbles in one pit. “But sir, how did you know?” “Because two of the sites were outside the indicator fan and should have had no granitic clasts”. “Oh!” We all went back together.



Figure 8. Indicator fans in Vermont, New Hampshire and southern Quebec: 1. Barre, 2. Braintree, 3. Brocklebank, 4. Knox Mtn., 5. Lebanon, 6. Glover, 7. Ascutney, 8. Cuttingsville, 9. Mt. Herford; solid black, igneous source rocks; black lines downglacier are 10% isopleths showing per cent of source rocks in till samples (Sources: 1, Larsen, 1972; 2-8, Great Pebble Campaigns at Norwich University, 1974-78 and 1980-81; 9, McDonald, 1967).

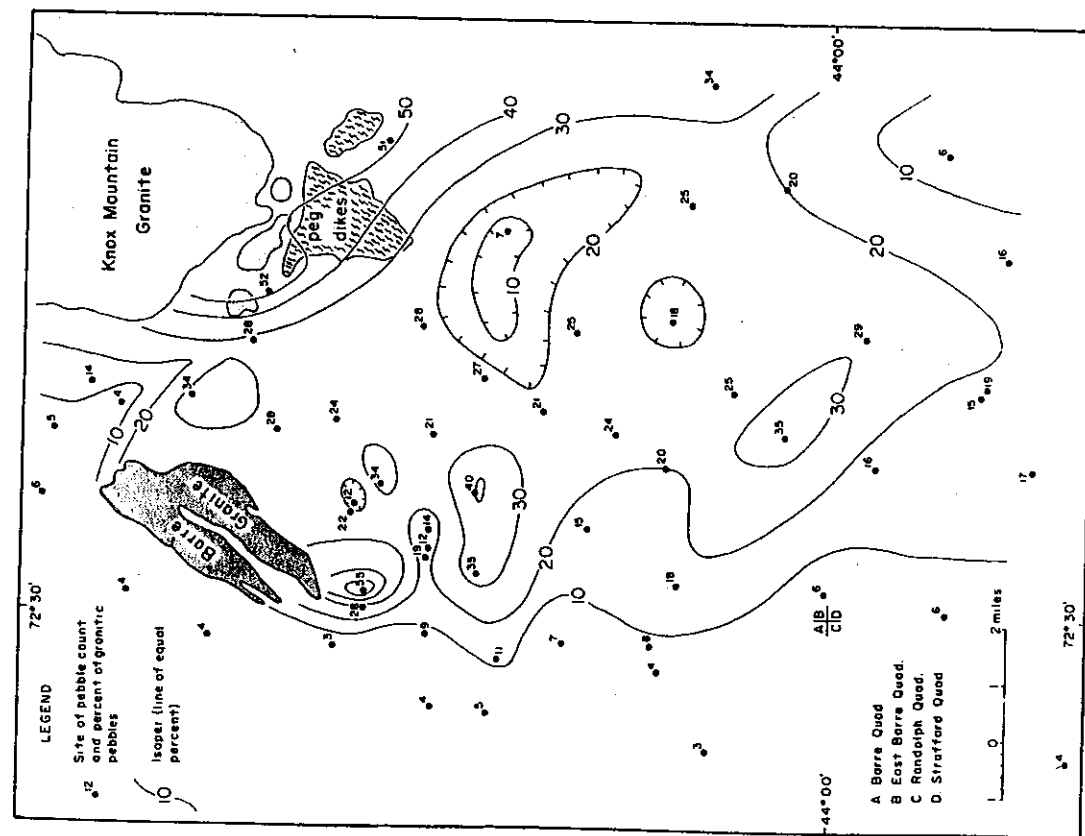


Figure 7. Indicator fan of pebbles from the Barre Granite pluton (Larsen, 1972).

Precambrian augen gneiss erratics in central Vermont (Larsen)

In their "Terranes of Northfield, Vermont", Richardson and Camp (1918) reported on the presence of augen gneiss boulders up to ten feet in diameter three miles west of Northfield on Rocky Brook and near the base of Burnt Mountain. They also reported augen gneiss erratics in the adjacent town of Roxbury. I did not pay attention to these publications until Diane Vanecek, Norwich '79, during the course of a senior mapping project, discovered an augen gneiss erratic 1.3 mi west-northwest of Northfield village. We collected the 1.3-ft boulder because it was unique, placed it on a dolly in the lab and named it the "Great Pebble" as an icon for the Great Pebble Campaigns. The rock was characterized by classic augen, light brown eye-shaped feldspars with reaction rims, and an irregular blotchy pattern of dark-colored ferromagnesian minerals. Over the years many erratics of augen gneiss have been identified in central and northwest Vermont (Fig. 9).

Curiosity finally took over and on June 8, 1981, I set forth in search of the source of the "Great Pebble". Using the S17°E direction of the Barre Granite indicator fan, I drew a line on the map trending N17°W from Northfield. Old Friends John Elson and Pierre LaSalle had sent me published maps showing augen gneiss plutons north of Montreal. As I traveled on the line I stopped every 10 to 15 miles at a till bank, gravel pit or gravel bar in a river. Nearly every time I stopped I eventually found a clast of augen gneiss similar to the "Great Pebble". A farmer in Franklin on the Quebec border had just cleared a forest and had bulldozed many Precambrian erratics into a big pile that included three rounded boulders of augen gneiss.

Five days later on a rainy, cold morning, I decided to make a run for it toward N17°W and by mid-afternoon I was at La Glaciere 68 miles north of Montreal and right on the Lac Croche pluton, a source of augen gneiss. It was a big disappointment. Although striations measured S19°E, feldspars in the weathered augen gneiss were bright pink instead of light brown as in the "Great Pebble". I pulled out the map that John Elson had sent to me and compared it to the topo sheets I had and realized that there was another possible pluton near St-Gabriel-de-Brandon 25 miles to the east. So I traveled northeast of St. Gabriel-de-Brandon on the St. Didace pluton and there it was in a roadside outcrop, the same lithology, color and appearance as in the "Great Pebble". The joy of discovery!

In an effort to pin down the source of the "Great Pebble", Nathan Donahue (1999), Norwich '01, and his advisor David Westerman did a geochemical analysis of the "Great Pebble" and three other augen gneiss erratics from central Vermont. They collected and analyzed three samples of bedrock from the Lac Croche pluton and six from the St. Didace pluton. On a plot of Niobium versus Yttrium, samples of Lac Croche and St. Didace bedrock fall in separate envelopes and the four samples of augen gneiss erratics fall within the St. Didace envelope. We are convinced that the "Great Pebble" and its siblings were derived from the St. Didace pluton and that the Laurentide ice sheet moved them due south in Quebec toward the Champlain valley and then southeast across the Green Mountains into central Vermont.

Additional evidence for southeast movement of the ice sheet in Vermont comes from the distribution of clasts of reddish-brown quartzite derived from the Monkton quartzite (Fig. 10).



Figure 10. Distribution of some erratics of reddish-brown quartzite (solid dots) derived from the Monkton quartzite (solid black areas).

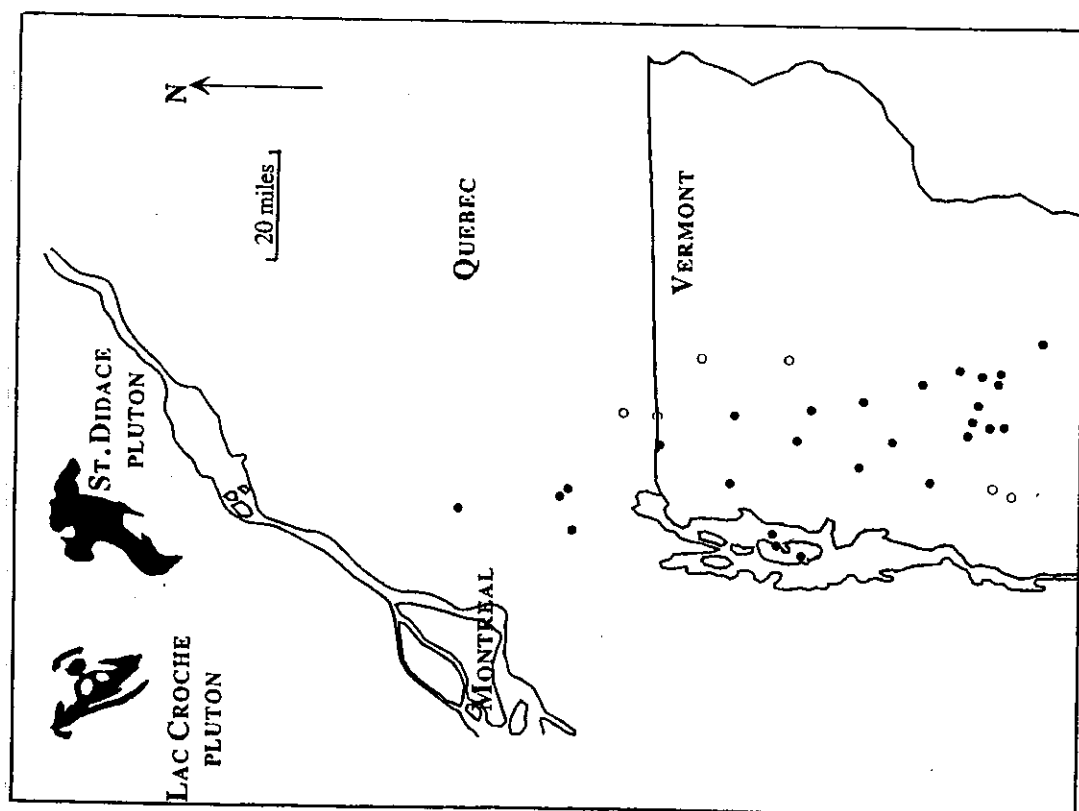


Figure 9. Map showing partial distribution of augen gneiss erratics in Vermont and Quebec. Open circles are locations where careful searches revealed no augen gneiss erratics. Erratics mapped by F. D. Larsen and N. P. Donahue.

GLACIAL DEPOSITS

Till: Till is the most common glacial deposit in central Vermont. It covers most upland areas with a blanket of poorly sorted rock debris that was deposited directly from the ice. The color and composition of unweathered till closely resemble those same characteristics in local bedrock, because most of the particles in the till have been transported only a short distance from their bedrock source. In both the Northfield quadrangle (Larsen, 1984) and the Montpelier quadrangle (Larsen, 1999a) the ice sheet had three main sources of material to incorporate into its base and recycle into deposits of till. They are: (1) the area of chlorite-rich rocks of the Moretown and Stowe Formations west of the Taconian Line (Fig. 1), (2) the area of calcareous quartzites and phyllites of the Waits River and Stowe Formations east of the Taconian Line, and (3) unconsolidated silt and clay in the valley bottoms of the North Branch (Montpelier quad) and the Dog River (Northfield quad). These latter deposits are believed to have formed in Glacial Lake Merwin, a proglacial lake that formed in the Winooski drainage basin during the advance of the Late-Wisconsinan ice sheet (Larsen, 1999b).

In the two above-mentioned quadrangles three facies of surface till are recognized on the basis of color and texture. Till derived from the Moretown and Stowe Formations are sandy to silty in texture and light to dark greenish gray in color when fresh. When weathered, the till is yellowish brown to dark orange brown because of magnetite and greenstone that are easily altered to dark iron oxides by acid rain percolating down through the upper soil zone. The maximum observed thickness of "Moretown-type" till is 30 ft (9.1 m) at a borrow pit on Culver Hill Road 0.2 mi, N30°W of Wrightsville Dam.

Till derived from the Waits River and Northfield Formations is gray in color when fresh and light brown when weathered. The lodgment till is sandy to silty and very compact due to cementation by calcite derived from clasts of calcareous quartzite. The ablation till is loose, sandy and yellowish brown in color. Large brown erratics in the till are calcareous quartzite from which soluble calcite and ankerite have been removed by ground water, leaving an iron-stained rock. Sometimes when all of the soluble minerals have been removed, only pockets of loose brown sand are left behind as phantoms.

In the valley bottoms of the North Branch, Dog River and Mad River, the till has a definite "lacustrine component" of gray silt and clay. At a site 0.85 mi (1.4 km), N38°E of the Vermont State Capitol the till is a mixture of two distinct sediments, gray, thinly-bedded varves of silt and clay and yellowish-brown gritty Moretown-type till with numerous pebbles and cobbles, some of which are phantoms derived from greenstone. Slabs of deformed varves and irregular masses of till have been placed adjacent to each other by shoving at the base of the ice sheet. It is difficult to tell whether the till at this particular site was deposited during the Middlesex readvance (see Fig. 3) or earlier during the Late-Wisconsinan ice advance. In either case the till was deposited and then was covered with over 100 ft of undeformed varved silt and clay deposited in Lake Winooski and subsequently exhumed by stream erosion by the North Branch.

Ice-contact deposits: The most significant ice-contact deposits in central Vermont are esker sediments and associated proximal and distal subaqueous outwash. Larsen (1987b) developed a model stratigraphic section for gravel deposits in the north-draining Dog River valley that is similar to one presented by Rust and Romanelli (1975) for the Ottawa, Ontario, region. The model stratigraphic section is applicable to deposits in the south-draining Second Branch valley (Stop 1) and the north-draining Stevens Branch valley (Stop 4). The model (Fig. 11) is comprised of 3 fining-upward units that are often capped by stream-terrace deposits.

Unit 1 consists of poorly sorted pebble gravel with cobbles and boulders in south-dipping cross beds up to 10 ft high. The deposits are confined to narrow elongate areas that are interpreted to be segments of eskers formed by subglacial meltwater streams. The sediments are rarely collapsed except on their outside margins. In several localities in the Dog River valley esker sediments rest directly on meltwater-scoured bedrock. The model of subglacial fluvial erosion described by Gustavson and Boothroyd (1982) for the Malaspina Glacier, Alaska, appears to be applicable here. In their model a groundwater table or potentiometric surface rises upglacier from a proglacial lake and supplies the head required to drive meltwater and meteoric precipitation through a subglacial stream system.

Overlying the coarse-grained esker deposits are 3 to 15 ft, or more, of medium to very coarse sand and pebble gravel in trough-cross bed sets 0.5 to 2 ft thick. The dip direction of cross beds generally is away from the glacier front, but occasionally at right angles to the axis of the esker. At some localities large channel deposits truncate well-bedded medium to fine sand. The channel deposits contain structureless medium sand and slabs of bedded fine sand and silt up to 1.0 ft in length. The observations indicate that Unit-2 sediments were deposited by density currents close to the mouth of a subglacial tunnel. Unit 2 sediments are characterized as proximal subaqueous outwash.

Unit 3 deposits consist of fine to very fine sand, silt and clay that are often rhythmically bedded. In places, angular ice-rafted clasts are common. Ripple-drift cross-lamination formed by south-flowing turbidity currents indicates formation of Unit 3 as distal subaqueous outwash. Unlike Units 1 and 2, which are relatively undeformed except for minor faults, Unit 3 displays various degrees of deformation from highly deformed at the base of a section to less deformed at the top. This indicates that lake-bottom sedimentation was contemporaneous with the melting of buried ice.

LATE-GLACIAL DEPOSITS

Basically, the term *late glacial* as used here refers to the time when glacial lakes like Lake Winooski and Lake Hitchcock occupied the valleys of central Vermont (Fig. 2). The next time unit, *postglacial*, started when the glacial lakes drained from the area and stream erosion and deposition commenced. Lake Winooski drained when the ice margin blocking the lower Winooski valley retreated in the Jonesville-Bolton area and Lake Hitchcock

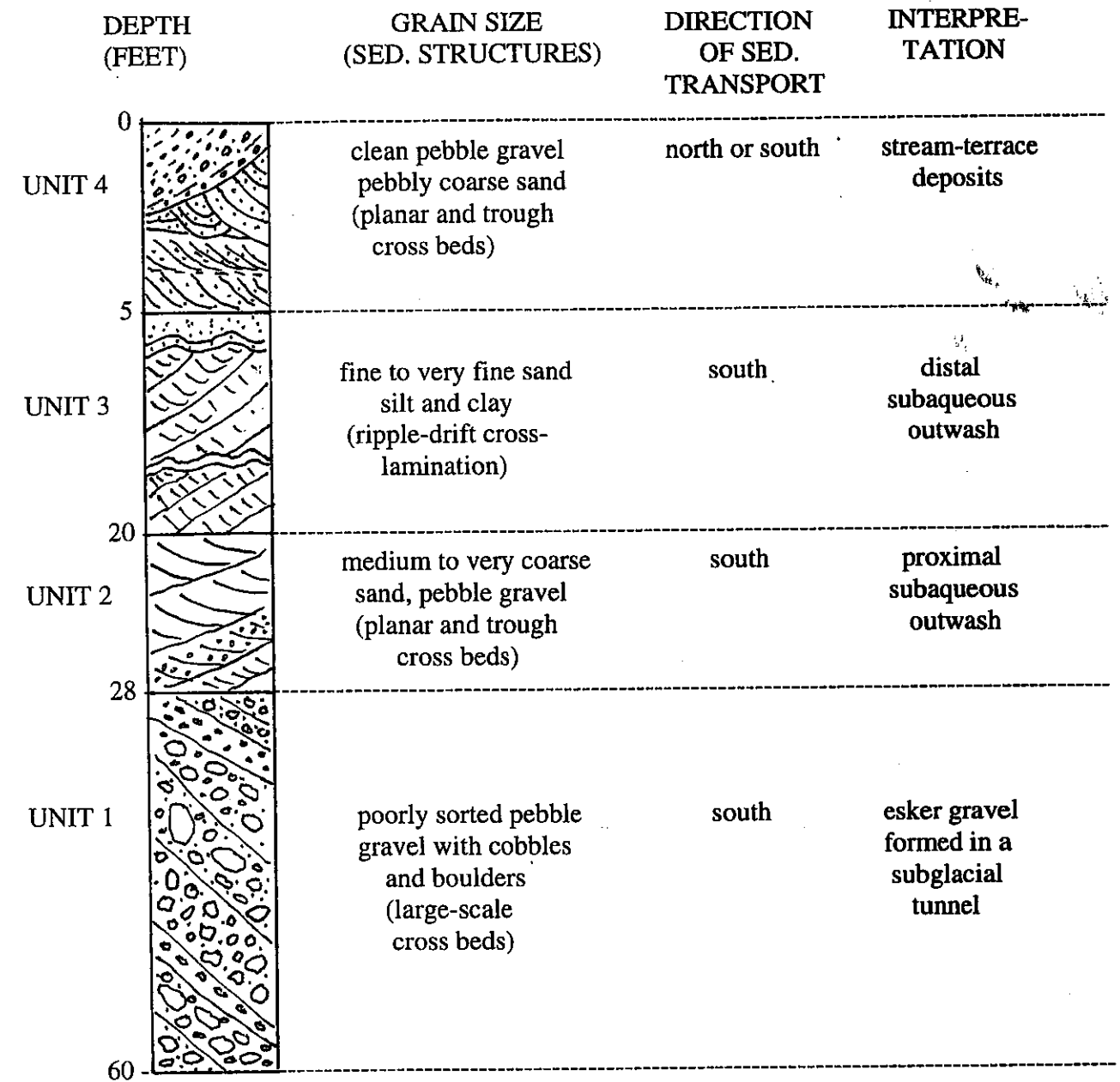


Figure 11. Generalized stratigraphic section for gravel pits in the valleys of the Dog River, Second Branch of the White River and the Stevens Branch of the Winooski River. Units 1, 2 and 3 constitute a fining-upward sequence produced during northward retreat of an ice margin in a lake. Sediment is delivered to the valley floor by south-flowing turbidity currents issuing from a subglacial tunnel. Unit 4 was formed after the lake drained (Compare with Rust and Romanelli, 1975; depth scale is not linear).

drained when a drift dam at Rocky Hill, CT was breached. In each case, postglacial deposition started when the glacial lake drained.

Lake-bottom deposits: A varve represents an annual glacial-lake deposit consisting of silt or very fine sand deposited during the summer when the lake was ice-free and a thin clay layer formed during the winter when the lake was frozen over. Layers of uncollapsed fine to very fine sand and clay-silt varves commonly are found at lower elevations in the valleys of the Winooski and White Rivers. Excellent exposures of classic clay-silt varves deposited in Lake Winooski will be viewed at Stops 8 (Wrightsville Dam) and 10 (Waterbury Reservoir). Detailed measured sections (varve curves) for those two stops plus a varve section at Muzzy Road 4.2 mi, S40°W of the Vermont State Capitol have been correlated with Antevs' (1928) New England varve curve for the Winooski valley, and are discussed below under the descriptions for these stops (Larsen et al, 2001).

Unique thick varves, some greater than 3.3 ft in thickness, occupy the valley of the Third Branch of the White River at and south of Randolph and in Ayers Brook valley 1.0 mi north of Randolph. The thick varves are well exposed at numerous landslide scars along the Third Branch. In 1978, David Brown, Norwich '79, measured a section 2.3 mi southeast of Randolph village that was 46 ft high. The section is composed of 13 varves with an average thickness of 3.5 ft. The winter portion of a "thick varve" has up to 2 in of silt and clay and the remaining portion of the varve has layers of fine to very fine sand and silt that display a variety of sedimentary structures. Summer layers in the lower half of the section are characterized by evenly spaced flat laminations. In the middle of the summer layers, laminations pass upward into sets of sinusoidal ripples, whose crests are locally offset to the south-southeast, the inferred current direction during sedimentation. In the upper half of the section A- and B-type ripple cross bedding becomes more common and also indicates transport to the south-southeast. The above description is typical for many other sections.

Not all the bottom deposits of Lake Hitchcock in the Randolph area consist of thick varves. Throughout those areas of Lake Hitchcock in the upper Ayers Brook valley and upper Third Branch valley, thin clay-silt varves occur in stream channel and bank exposures. The thick varves described above are unusual in that the summer layers are so thick, often over 3 ft in thickness. We attribute the great volume of sediment in thick varves in the Randolph quadrangle to rapid retreat of a stagnant ice margin with the production of great quantities of meltwater, and to the focusing of deposition of sediment in a narrow valley.

Thick lake-bottom deposits were examined on the north bank of Ayers Brook about 0.75 mi north of Randolph village. Thick varves as sedimentation units were not as distinct as at other locations described above. Two lines of evidence suggest that sediments in the section were formed by north-flowing turbidity currents. First, there is the overall dip of about 6° to the north of the layers and second, the presence of numerous deformational structures such as north-verging slump folds, north-tilted flame structures and ball-and-pillow structures stacked to the north. Thick varves only extend up the Ayers Brook valley for about a mile north of Randolph. It appears that the thick varves are part of a large

subaqueous fan that was built by turbidity currents flowing east in the upper Third Branch valley. At the site of Randolph village the currents spread out and flowed north up the Ayers Brook valley and east and south-southeast in the lower Third Branch valley. If such a fan existed, it would create a high or a bulge in lake-bottom sediments at Randolph between Ayers Brook valley and the lower Third Branch valley. This would explain the depression in which Holocene sediments were deposited in Ayers Brook valley (see "Ponded sediments of Ayers Brook valley" below).

Deltaic deposits: As mentioned above, there are as many as 15 or more deltaic landforms on the projected shoreline of Lake Winooski (Fig. 2), however, at only three has the topset/foreset contact been exposed. Excavating the topset/foreset contact and measuring the elevation in these Lake Winooski deltas to determine the amount of tilt due to rebound remains a project for the future.

Deltaic deposits of Lake Hitchcock have been observed at several sites in the Randolph quadrangle. Trimmed foreset beds have been observed at two localities. At one of these sites, the senior author (Larsen, 1987a, Stop 6, Lower Branch "delta") described a presumed topset/foreset contact at 749 ft ASL that is 23 ft below the projected level of Lake Hitchcock of Koteff and Larsen (1989). It was assumed that this so-called "delta" represented the level of a "low" Lake Hitchcock and a projected shoreline was drawn using 4.74 ft/mi to N21.5°W from the 749 ft elevation at Lower Branch delta. This line, when drawn in the Randolph quadrangle, turned out not to represent a "low" Lake Hitchcock shoreline, but to identify localities where flat post-lake fluvial deposits rest directly on flat lake-bottom deposits of Lake Hitchcock. These flat fluvial deposits represent the first sediments deposited directly on the bottom deposits of Lake Hitchcock when it drained. They have been named "terrace/fan deposits" (see below) in the Montpelier quadrangle (Larsen, 1999a), and inconsistently "fan/terrace deposits" in the Randolph quadrangle (Larsen et al, 2003a).

At this time, there are no identified Lake Hitchcock deltas in the area shown on Figure 2. It is possible that Lake Hitchcock drained while the ice margin was in this area and any deltas constructed were small and have been removed by erosion.

POSTGLACIAL DEPOSITS

Terrace/fan deposits: Flat to gently dipping layers of pebbly medium to coarse sand, pebble gravel with cobbles are found lying directly on flat lake-bottom deposits of Lake Winooski in the Montpelier quadrangle and of Lake Hitchcock in the Randolph quadrangle. The topographic expression where terrace/fan deposits are found consists of flat stream terraces and gently sloping alluvial fan surfaces that grade laterally into each other, and are elevated above the modern stream system. These landforms are the remnants of once-larger fluvial deposits that were laterally extensive, and have been eroded into many separate segments by the stream systems that formed them.

An important example 0.5 mi southeast of downtown Montpelier is the flat surface underlying College Street and Vermont College (now Union Institute and University). The surface is underlain by 10 ft of pebble gravel and pebbly sand of fluvial origin resting on flat lake-bottom sand. This surface was once under 300 ft of water of Lake Winooski. In the Randolph quadrangle the southeast-sloping surface on which Gifford Memorial Hospital is located is underlain by terrace/fan deposits. Pebbly sand in southeast-dipping trough cross beds were observed in a house excavation 5 to 6 ft deep. A similar surface sloping to the northeast is located one mile to the north. These two surfaces represent the fluvial cover on the north and south segments of the subaqueous fan described above. The central portion of the fluvial-capped subaqueous fan ("downtown Randolph") has been removed by erosion by the Third Branch.

"Ponded sediments of Ayers Brook valley": Organic silt and fine sand interbedded with peat-like material containing wood fragments occur on the Hodgdon farm 1.20 and 1.24 mi, N18°E of the junction of Vermont Routes 12 and 12A in Randolph village (Larsen et al, 2003a). The ponded sediments display 3 separate units, which from the bottom up are: (1) more than 6 ft of dark gray organic silt with wood fragments, (2) 2.5 ft of interbedded peaty organic material, silt and fine sand with small-scale cross beds formed by ripples and (3) 3.5 ft of interbedded silt, very fine sand, and fine sand with ripple cross beds and relatively little organic material. The ponded sediments are unlike lake-bottom deposits of Lake Hitchcock. In 2001, dating of peat-like material gave an age of 8700 ± 150 ¹⁴C years BP (Geochron No. GX-28660). Professor Lauren Howard of Norwich University (pers. commun., 2002) reports that the wood samples that he was given were "all hardwood and mostly *Liriodendron tulipifera* (tulip poplar). It must have been a warmer time as tulip tree rarely comes up into southern Vermont today, and is mostly found from New York south".

In an effort to confirm the 8700 BP date, two more samples of wood were collected for dating in the fall of 2002 by Larsen, Dunn and Springston (2003b). A sample of wood thought to be *Fraxinus* (ash) (L. Howard, pers. commun., 2003) was collected 2.6 ft below the 8700 BP sample site and has an age of 8910 ± 120 ¹⁴C years BP (Geochron No. GX-29693). The second sample, a small log, was collected from the base of a small gully 1.73 mi, N10°E of the first sample site. The age is 9980 ± 130 ¹⁴C year BP (Geochron No. GX-29692) and has been identified as *Picea* (spruce) (L. Howard, pers. commun., 2003).

A test boring by the Vermont Agency of Transportation and logged by Alan Randall (FOP, USGS retired) was taken at the Peth Road bridge over Ayers Brook in 1985. The bridge is 0.6 mi south of the 9980 BP sample site. Randall reports, "10-13 ft, black (organic) to gray silty clay; 13-44 ft gray silt". The dark gray organic silt is exposed in a logging yard on the east side of VT Rt 12 0.2 mi south of Peth Road and in several nearby auger holes.

When Lake Hitchcock drained there was still a small body of water in Ayers Brook valley that was dammed up by thick lake-bottom deposits in the form of a subaqueous fan that stretched across the lower Ayers Brook valley. This body of water has been named Ayers Pond (Larsen, 2002) and initially could have been 3.0 mi long and 0.5 mi wide. Ayers Pond existed over 1280 carbon-14 years in the Early Holocene and collected many different

kinds of organic material were deposited there. According to Davis and others (1980), the climate in New England at 9000 BP was 2°C warmer than at present. This would explain why *Liriodendron* was found in the Ayers Pond sediments of central Vermont north of its present-day range.

Stream-terrace deposits : Sand of all sizes, pebbly sand, pebble gravel and pebble gravel with cobbles are common constituents of stream terraces throughout central Vermont. The stream-terrace deposits display small- and medium-scale cross beds formed by the migration of ripples and dunes, respectively. Stream terraces occur below terrace/fan deposits and above the modern flood plain. Only a few low stream terraces have been mapped in the Montpelier and Barre West quadrangles, but terraces at several levels are common in the Northfield and Randolph quadrangles. In the Randolph quadrangle there are many terraces where fining-upward sequences formed by meandering streams have been exposed. A neat problem to be solved in the future concerning terraces in the Randolph quadrangle relates to the apparent slow lateral migration of the Third Branch across a flood plain and the episodic rapid down cutting, abandonment of the flood plain as a terrace, and a repetition of the process.

Alluvium: Flood-plain deposits share the same grain sizes described above for stream-terrace deposits.

GLACIAL, LATE-GLACIAL AND POSTGLACIAL HISTORY

The direction of ice movement in central Vermont was between south-southwest to southeast. Striation direction varies greatly in any one quadrangle. In the Dog River valley striations follow the valley bottom to the south-southwest. On the crest of the Worcester Range north of Montpelier and on the Northfield Mountains striations trend to the southeast. In the lee of the Worcester Range in the Montpelier quadrangle ice direction was close to due south. Indicator clasts were brought to central Vermont from a sector between north and northwest. The Late Wisconsin Laurentide ice sheet that covered all of Vermont reached a maximum extent on or near Long Island about 21,750 ¹⁴C years ago, and then the ice began to recede from end moraines before 19,000 ¹⁴C years ago (Sirkin, 1982). When the climate turned warmer the ice sheet thinned and the ice margin retreated to the north, slowly at first then more rapidly in northern New England (Ridge et al, 1999).

During northward of the ice sheet in the Connecticut River valley, the ice margin was accompanied by a northward-expanding glacial Lake Hitchcock that was dammed up at Rocky Hill, Connecticut, and had an outlet in New Britain, Connecticut. Eventually with further retreat of the ice, Lake Hitchcock extended north into the First, Second and Third Branches of the White River and into Ayers Brook valley. Great quantities of fine-grained lake-bottom sediment were continually deposited in the Third Branch valley near Randolph as the ice margin retreated northward into the Roxbury quadrangle. As soon as the ice margin in the upper Third Branch valley retreated north into the drainage basin of the north-flowing Dog River, fine-grained sediment carried by meltwater from the ice would have been deposited in glacial Lake Roxbury with a threshold at 1,010 ft ASL at Roxbury. This

essentially cut off the major supply of fine-grained sediment to Lake Hitchcock by way of the Third Branch. However, meltwater from the ice margin still flowed through Lake Roxbury, over the threshold and down the Third Branch, but with much less suspended sediment. When the ice margin in the Dog River valley retreated northward, Lake Roxbury dropped about 80 ft to the level of Lake Winooski. At that time, water from Lake Roxbury ceased flowing down the Third Branch and the Roxbury threshold was abandoned.

Meanwhile near East Randolph in the valley of the Second Branch, subglacial streams flowing under stagnant ice formed eskers on the floor of the valley. As the ice margin retreated, the eskers were covered with lake-bottom deposits formed in Lake Hitchcock. The waters of Lake Hitchcock extended north into Williamstown Gulf. The ice margin retreated north past a 915-foot threshold into the drainage basin of the north-draining Stevens Branch cutting off the supply of fine-grained sediment to Lake Hitchcock. Subglacial streams formed a series of large eskers in the Stevens Branch valley and meltwater flowed through glacial Lake Williamstown, over the Williamstown threshold and down through Williamstown Gulf into Lake Hitchcock.

When Lake Roxbury dropped 80 ft to the level of Lake Williamstown that was the beginning of glacial Lake Winooski, which also drained down through Williamstown Gulf. As long as the retreating ice margin dammed the Winooski valley and prevented meltwater from flowing to the Champlain basin, Lake Winooski existed. When the ice margin was in the lower Winooski valley in the vicinity of Bolton and Jonesville, Lake Winooski drained and a lower lake, glacial Lake Mansfield was formed with a threshold southwest of Huntington. At that time, the Williamstown threshold was abandoned.

According to varve correlation, Lake Winooski lasted a minimum of 312 years. During those 312 (or more) years Lake Hitchcock drained. Large elongate fluvial gravel bars with well-sorted pebble gravel were formed in the upper Second Branch valley below the level of Lake Hitchcock. This indicates that Lake Hitchcock lowered while meltwater from Lake Winooski flowed down through Williamstown Gulf forming the gravel bars. The draining of Lake Hitchcock marked two important events in the Randolph quadrangle: (1) local streams all poured sediment out onto the exposed lake bottom to form terrace/fan deposits and (2) the entrapment of water behind a sediment dam in the Ayers Brook valley and the beginning of sedimentation in Ayers Pond, which eventually received organic material during the Early Holocene.

There are several localities in four different quadrangles in central Vermont where till overlies clay-silt varves that were deposited either before or after the Late Wisconsinan ice advance. At one locality near the center of the Montpelier quadrangle, the ice sheet readvanced and picked up a 6-ft long mass of fine-grained sediments with organic material dated at 11,900 ^{14}C years BP. That particular site clearly documents readvance after the Late Wisconsinan ice advance. How many of the other till over varves sites are related to the Middlesex readvance at 11,900 ^{14}C years BP, or to the Late Wisconsinan ice advance, we simply do not know.

The draining of Lake Winooski occurred not more than 312-plus years after the draining of Lake Hitchcock, and resulted in a set of terrace/fan deposits in the Winooski drainage basin. Subsequent downcutting by the Winooski River and its tributaries resulted in the entrenchment of the stream system into lake-bottom deposits. Lateral migration of the streams followed by downcutting produced stream terraces at several levels. One or more of those terraces could be on fluvial grade to Lake Mansfield to the west of Montpelier.

Rebound of the land due to the removal of the weight of the ice did not occur until after Lake Hitchcock drained. We know this because the tilted water plane of Lake Hitchcock measured by Koteff and Larsen (1989) is planar, and not curved. If rebound of the New Britain threshold had occurred before the lake drained causing the lake level to rise, then the youngest deltas in the north would have been formed at a higher lake level, thereby producing a curved (concave-up) water plane. Sediments in Lake Champlain changed from saline Champlain Sea deposits to fresh-water lake deposits about 10,000 years ago (Hunt, 1980). This indicates that the threshold for Lake Champlain, now at 95 ft ASL, was uplifted to sea level about 10,000 BP. Since total rebound in the upper Champlain valley is about 400 ft, about 75% of rebound occurred after Lake Hitchcock drained at 13.7 ^{14}C ka (Ridge et al, 1999) and 10 ^{14}C ka. Parent and Occhietti (1999) show a curved water plane for glacial Lake Memphremagog in southern Quebec, which indicates that rebound had started while the retreating Laurentide ice margin was located in southern Quebec. We conclude that uplift of the Vermont landscape was rapid for about 4,000 years following the draining of Lake Hitchcock, and then it slowed during the Early Holocene.

Today, the Winooski and White Rivers and their tributaries have cut down to bedrock in many places, thus slowing downcutting and increasing lateral cutting in easily eroded lake-bottom sediments. Hurricanes and thunderstorms produce torrential downpours that result in massive slumps and landslides. In June, 1998, the Third Branch experienced a greater than 100-year flood based on a 58-year stream record at the Ayers Brook gaging station. The town of Randolph alone experienced damage greater than \$1.4 million to roads and infrastructure.

ACKNOWLEDGEMENTS

All the acknowledgements in our published NEIGC field trip articles are still extant. We thank landowners for permission to study on their private property. We thank Norwich University, the University of Vermont, the Vermont Geological Survey and U.S. Geological Survey for logistical and financial support. Many thanks to: Nathan Donahue for shopping, driving the lunch truck and cleaning the exposures, Lori Barg for her assistance in the field, UVM grad students Michael Eller and Luke Reusser for driving the vans and preparing the exposures; we thank Professor Lauren Howard, chairman of the Norwich University Biology Department for identification of Ayers Pond wood; we thank the staff of the Little River State Park for facilitating geologic study in the Waterbury Reservoir.

ROAD LOG - DAY 1

Assembly time and place: Saturday, May 24, 2003, 8:00 AM. South parking lot at Econo Lodge, Montpelier, Vermont. Cumulative mileage will be given, including average mileage in pits. Stops on private property require permission from the land owner.

0.0 mi Proceed west from Econo Lodge on Derby Drive, north on Mountainview and northwest on National Life Drive; at 1.0 mi red light, turn left on Memorial Drive; stay in left lane and at 1.5 mi pass under I-89 and proceed south. At 2.3 mi is important geologic contact on the northbound lane known as the Taconian Line or Richardson Memorial Contact. The TL/RMC is recognized by the abrupt change from light greenish-gray rocks of the Moretown Formation to dark gray slate of the Northfield Formation. All bedrock observed beyond this site to Stop 7 is in the Waits River Formation. This formation is made up of thin to thick layers of calcareous quartzite and phyllite. The quartzite formed originally as turbidites that weather to a chocolate brown when the calcite is leached out. When the calcite is redeposited cementation commonly occurs.

24.5 mi Leave I-89 at Exit 4, proceed east then north and northeast on curvy VT Rt 66 to VT Rt. 14 at East Randolph; At 30.3 mi turn right on VT Rt 14, proceed south over terraces carved in Lake Hitchcock bottom deposits overlying esker gravels; At 32.2 mi turn left into pit and circle the wagons.

STOP 1: PIT OWNED BY LARRY PICKETT

Figure 12 shows the location of some of the better exposures. CAUTION is required near overhanging exposures. Scraping with a tool is frustrating because of the presence of many cemented layers. This is a problem in "Waits River terrain". The bulk of the esker gravel mined out at this site went into the construction of the interstate highway, I-89, in the late 1960's. The purpose of this stop is to document the presence of (1) subglacial esker gravel and (2) proximal and (3) distal subaqueous deposits of Lake Hitchcock and the conditions under which they formed. The stratigraphic sequence is similar to that described by Larsen (1987) for gravel and sand deposits in the Dog River valley. They are similar in that the esker gravel and proximal subaqueous outwash are relatively undisturbed, but the distal subaqueous outwash (varves) often are highly collapsed due to the melting of buried ice. This implies that the mouth of the subglacial tunnel was located in a deep reentrant in the ice margin with tongues of buried ice on either side. Cross beds in all three units dip to the south indicating a melt water source to the north from a northward-retreating ice margin during deglaciation. The projected level of Lake Hitchcock at this site is 748 ft ASL and the elevation of Lake Hitchcock bottom deposits extends up to an elevation of 680 f ASL just to the west of Stop 1. The floor of the pit is about 550 ft ASL.

Leave pit, turn right, proceed north on VT Rt 14; At 35.0 center of East Randolph; At 35.5 long pit on left, esker gravel mostly mined out, proximal and distal lake-bottom deposits

BRIEF DESCRIPTION OF SITES LABELED ON MAP

At H collapsed medium-thick varves dip 20° toward N33°W, difficult to study because of cemented layers, reverse faults.

At G a small remnant of unit #4, stream-terrace pebbly sand.

At F 2-foot thick varves are collapsed.

At E is location of 25-ft stratigraphic section showing units #1, #2 and #3.

At D at base of overhanging section is excellent exposure of unit #2 flat-bedded sand, proximal subaqueous outwash.

At C medium-thick varves contain many cemented layers.

At B grain size changes abruptly laterally and vertically, medium sand with dune cross beds overlie pebble gravel in south-dipping cross beds, alternate foresets have coarse pebble gravel with open-work structure interbedded with compact fine pebble gravel.

At A 42-foot stratigraphic section in esker gravel.

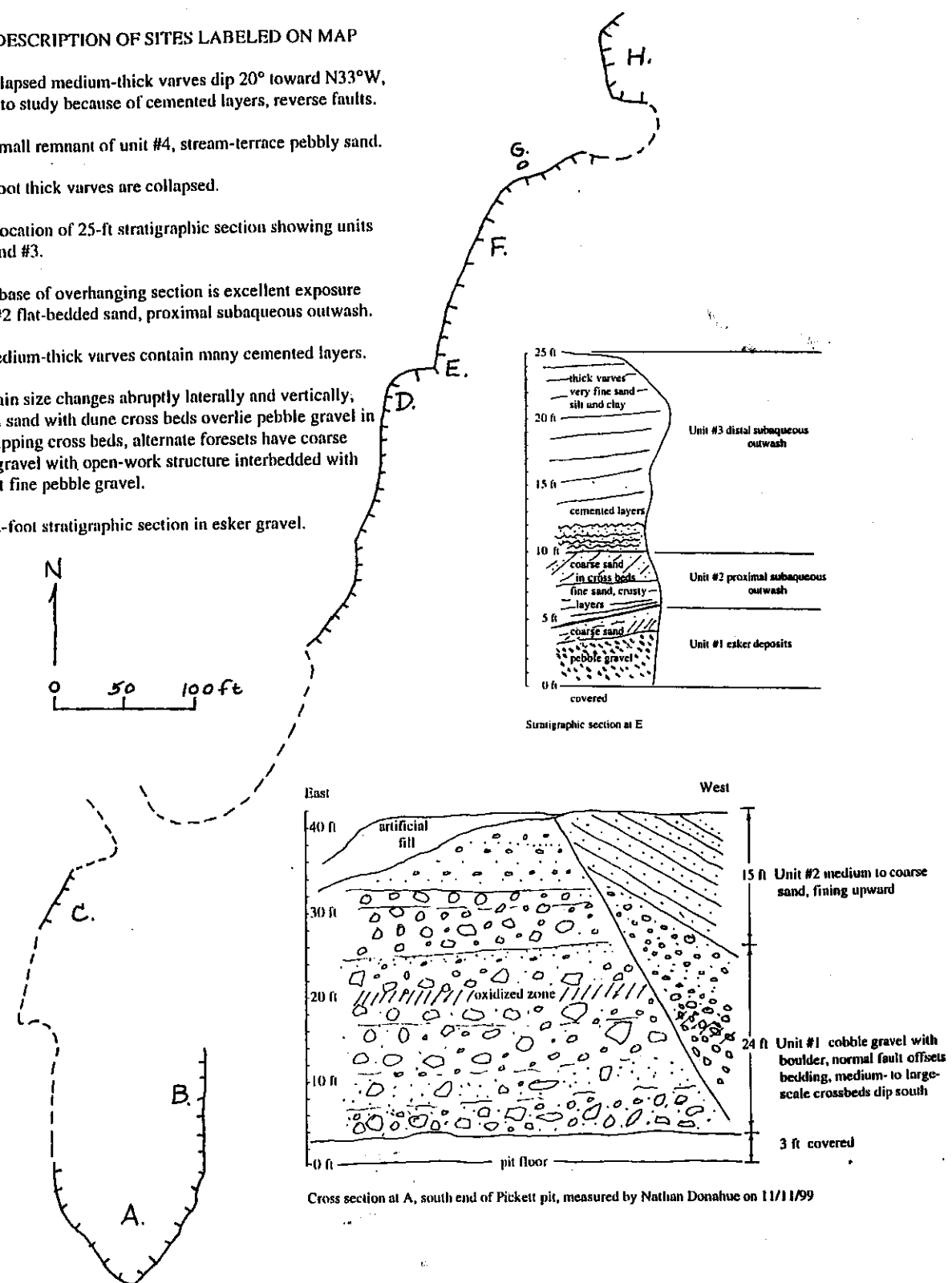


Figure 12. General outline of Larry Pickett pit 2.3 mi south of East Randolph, Vermont, on May 5, 2003. Better exposures are shown by hachured line, covered areas by dashed line. Exposures are subject to rapid change.

similar to those at Stop 1 are well exposed; at 36.1 driveway/farm road to Stop X on the right.

STOP X: STRATIGRAPHIC SECTION ON JOHN RACE PROPERTY

There is not room at this site for a large group. See John Race for permission to go on the land. A tight bend of the Second Branch is migrating south into bottom deposits of Lake Hitchcock (Fig. 13). The 35-foot section of distal subaqueous outwash was formed more than 100 ft below the surface of Lake Hitchcock with a projected level here of 760 ft ASL. Probably about 70 ft of Lake Hitchcock bottom deposits have been eroded from this site. Three separate packages of thin varves occur in a section dominated by layers of fine to very fine sand. A layer of clean pebble gravel in the middle of the section indicates a flood event.

Continue north on Rt 14; at 36.2 look back to the right for a view of the Stop X stratigraphic section. At 36.9 pit on left has some esker gravel, the proximal subaqueous outwash has many faults and cemented layers typical of gravel and sand derived from the Waits River Formation; At 37.0 pit on left, esker gravel mostly depleted, collapsed proximal subaqueous outwash; At 37.5 center of North Randolph.

38.0 to 38.4 travel on convex-up ridge, POSSIBLE HIGH GRAVEL BAR, cross Second Branch; At 39.1 closed pit on left; At 39.3 depleted pit on right in esker gravel; At 39.4 pit on right in north end of esker, cross beds dip south, faults.

At 39.9 note brick house on left, PIT TO WEST OF HOUSE IN LOW GRAVEL BARS was Stop 8 on 1987 NEIGC Trip A-3. The stop description is repeated here:

“WHEATLEY PIT IN GRAVEL BAR Pebble gravel with cobbles in flat beds was formed in the upper-flow regime The topography has been modified by man but inspection of aerial photographs of this area shows elongate landforms interpreted to be longitudinal bars formed by the outlet stream from Lake Winooski after Lake Hitchcock drained. This locality is 84 ft (25.6 m) below the minimum projected level of Lake Hitchcock” (Larsen, 1987).

40.0 Road to Wheatley pit on left, travel north on convex-up ridge (A HIGH BAR?)

40.1 Pit on valley floor ends

40.4 Sloping terrace on west valley wall was interpreted to be an ice-contact landform in 1987. In 1999, shallow holes revealed fine sand covered by pebble gravel. The fine sand is interpreted to be Lake Hitchcock bottom deposits and the pebble gravel to be a fan-terrace deposit (Larsen, 1999). The pebble gravel washed out onto the lake bottom when Lake Hitchcock drained.

40.9 Turn left on Old Post Road, travel slowly across CONVEX-UP GRAVEL BAR about 1000 ft long and 300 ft wide, property of Fred Locke; 41.2 turn left on VT Rt 14

41.9 Junction with VT Rt 65 in East Brookfield

42.1 Enter gravel pit on left in HIGH GRAVEL BAR, Stop 2

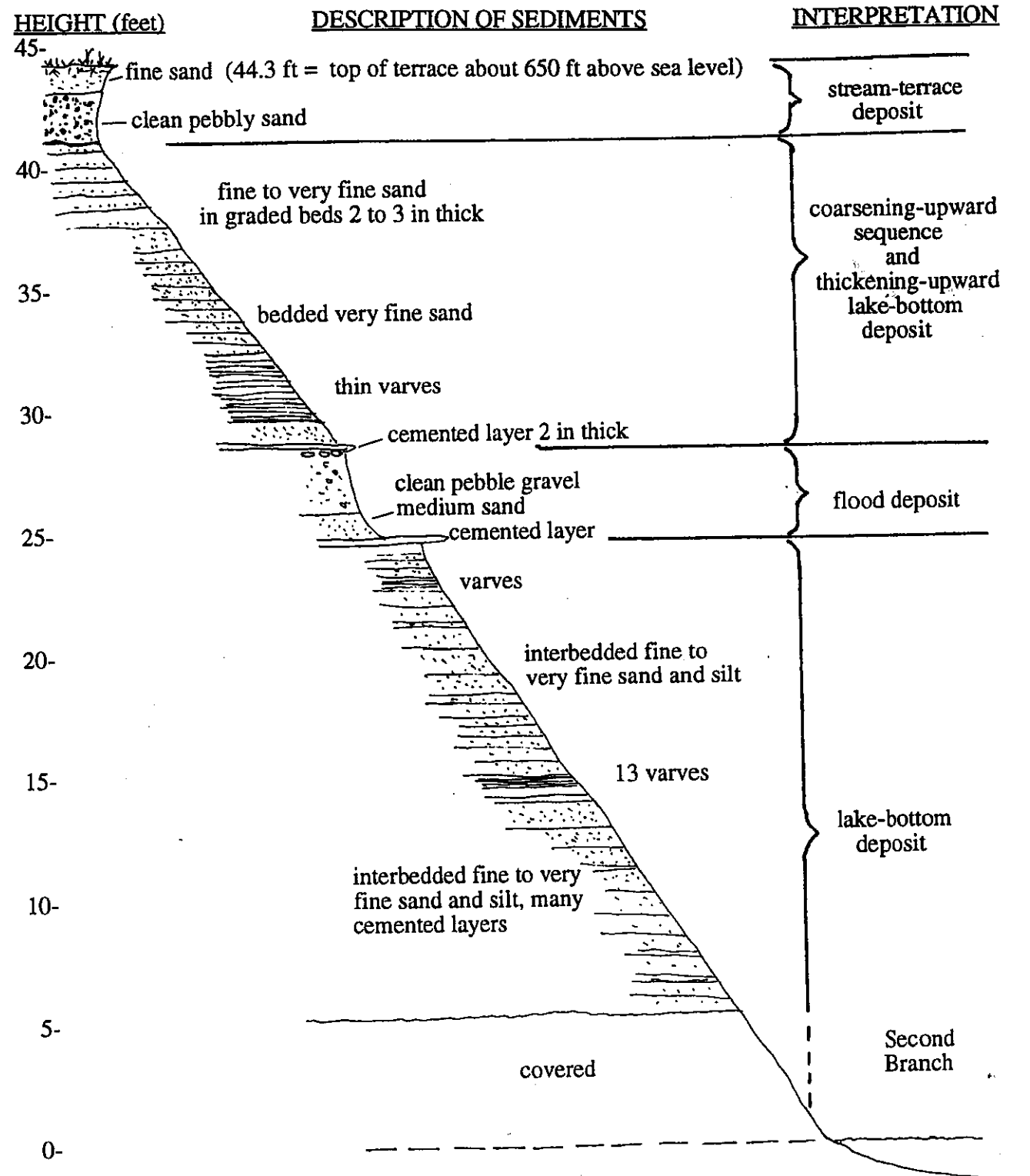


Figure 13. Stratigraphic section exposed in cutoff slope of the Second Branch of the White River, John Race property 1.25 mi south of North Randolph, Vermont

STOP 2: TOWN OF BROOKFIELD PIT (formerly John Wheatley pit) (Larsen)

It is not as obvious as it used to be, but this pit is located in a convex-up landform that is interpreted to be a high gravel bar. The surface elevation of the landform is about 730 ft ASL (from the topographic map) and the projected level of Lake Hitchcock is 785 ft ASL. The pit was opened in the mid-90's and was first observed and studied by the glacial geology class at Norwich University in the fall of 1999. The pit has 10 to 12 ft of pebble gravel with some pebble-cobble gravel at the base. Bedding is not apparent. Results of a single pebble count (Fig. 14) indicate that calcareous quartzite and phyllite from the Waits River Formation (Dwr) make up the bulk of the clasts in the pit. Working in pairs at three separate sites, each student carefully measured the strike and dip of five imbricated clasts and plotted poles to bedding on a stereonet (Fig. 14). Most of the flat clasts in this homogeneous gravel dip to the north and northeast suggesting fluvial transport from the northeast quadrant during time of deposition. It appears that there are several different levels of gravel bars in the upper Second Branch valley that extend 4.5 mi along the valley floor. Because the gravel bars occur well below the projected level of Lake Hitchcock, and presumably were formed by strong fluvial activity, it appears that they were formed by the outlet stream from Lake Winooski after Lake Hitchcock drained.

Proceed north on VT Rt. 14; At 43.0 GRAVEL BAR on right, on east side of valley floor.

43.4 alluvial fan on the right is a post-Lake Hitchcock feature formed by the dissection of a small "delta" located on the projected level of Lake Hitchcock

44.4 enter Williamstown Gulf, a V-shaped valley that was formed by outlet streams from proglacial lakes in the Winooski drainage basin during each advance and retreat of several ice sheets

46.2 CAUTION, TURN LEFT ACROSS TRAFFIC, ENTER DIRT ROAD, follow dirt road for 1.0 mi along north side of outlet channel from glacial Lake Winooski; At 47.2 park along side of dirt road for Stop 3

STOP 3: OUTLET OF GLACIAL LAKES WILLIAMSTOWN AND WINOOSKI (Larsen and Wright)

This is the current drainage divide between the north-flowing Stevens Branch of the Winooski River and the south-flowing Second Branch of the White River. Its current elevation 915 ft ASL (279 m) is the lowest drainage divide between the Winooski and White River drainage basins (Fig. 1). As soon as the margin of the ice sheet retreated north of this point, a lake was impounded, Glacial Lake Williamstown (Merwin, 1908; Larsen, 1972, 1987a), which expanded down the river valley as the ice front moved north. Shortly before the ice margin reached Montpelier, higher level lakes in the Dog River valley (Glacial Lake Roxbury) and the Mad River valley (Glacial Lake Granville) coalesced with Lake Williamstown to form Glacial Lake Winooski (Larsen, 1972, 1987b). The outlet to this much larger and expanding lake remained here at this drainage divide until the ice margin was located near Jonesville west of the Green Mountains and a lower threshold was deglaciated. Lake

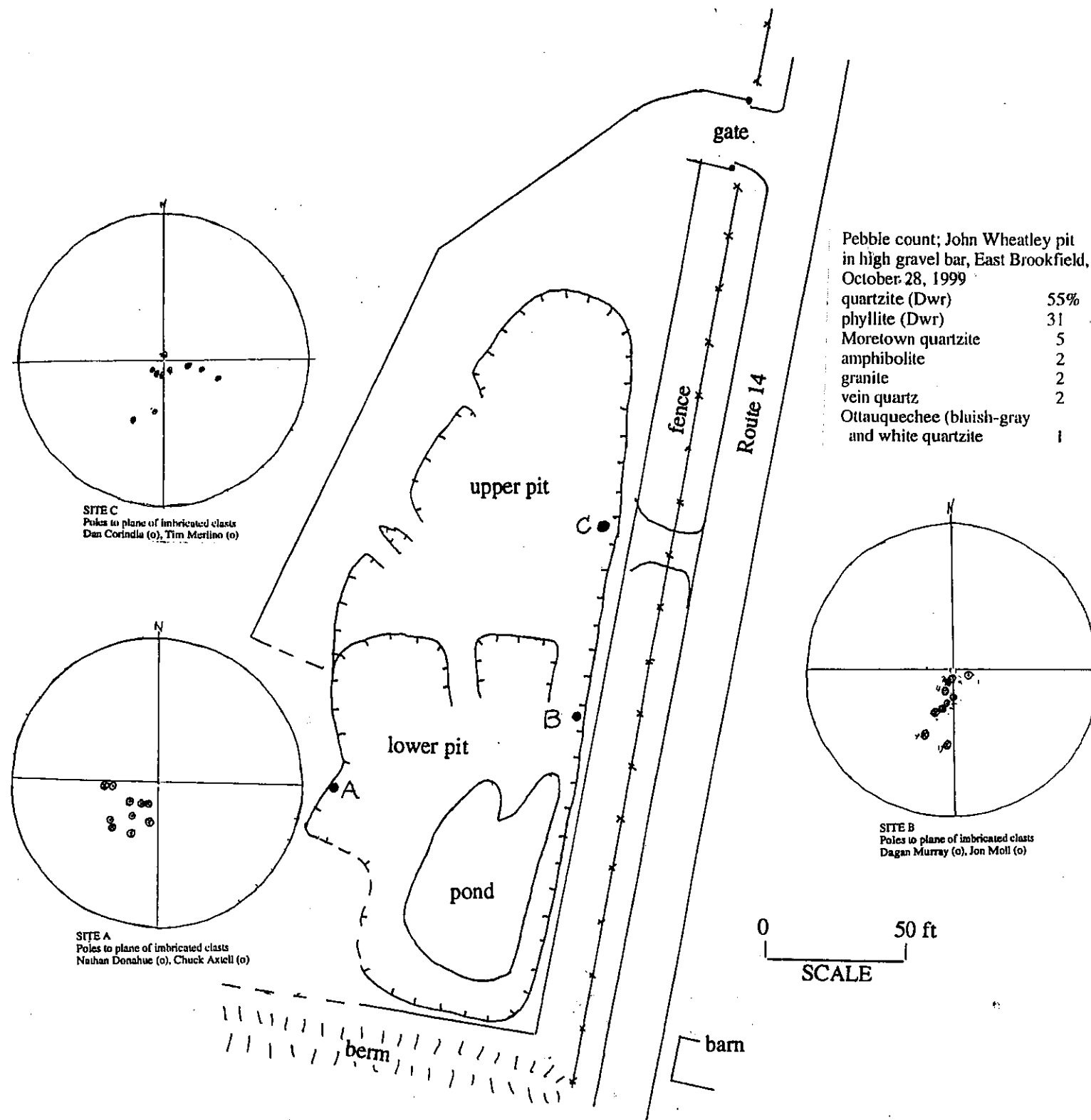


Figure 14. Pace and compass map of the Brookfield Town pit (formerly John Wheatley pit), East Brookfield, Vermont, by F.D. Larsen, 10/26/99

Winooski drained and a lower Glacial Lake Mansfield drained west into the Champlain valley over a threshold at 685 ft ASL (209 m) just southwest of Huntington. At that time, the outlet stream no longer flowed south and alluvial-fan sediments were deposited in the area of the drainage divide.

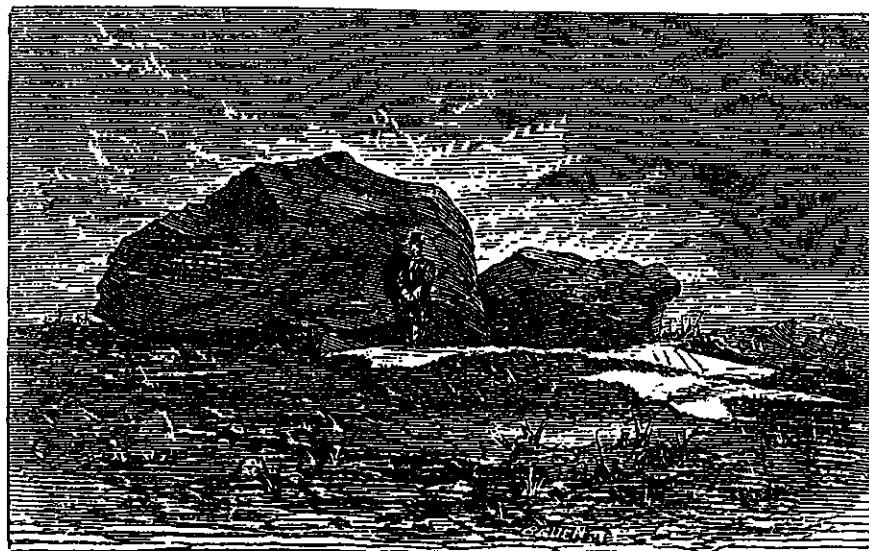
Proceed north on VT Rt 14; At 47.5 pit (now closed) on the left is in the Stevens Branch esker described by Wright (1999, Stop 2, p. 187).

49.7 Junction with VT Rt. 64 in the village of Williamstown

51.8 Barre Town and Williamstown pit, described by Wright (1999, Stop 3, p.188), is in the Stevens Branch esker

53.3 Turn right at pit entrance, Stop 4

FIG. 336.



Drift Striae on Mansfield.

Glacial erratic on top of Mount Mansfield. Striations shown on surface in front of small boulder (from Hitchcock *et al.*, 1861, p. 879).

STOP 4: STEVENS BRANCH ESKER (LAMSON PIT), SOUTH BARRE UTM Coordinates: 699080, 4893440 (Wright)

Note: Much of the following stop description has been taken from Wright (1999b).

Stevens Branch Esker

The Stevens Branch is a north-flowing branch of the Winooski River that joins the Jail Branch in Barre and joins the main stem of the Winooski River just east of Montpelier. Stewart and McClintock (1970) mapped most of the Stevens Branch valley as a kame terrace; the exceptions being an area mapped as a delta to the south and an area covered by lacustrine sediments to the north. My own mapping (Wright, 1999a) shows a different distribution of surficial materials in the valley consisting of two principal facies: (1) An esker facies generally consisting of coarse-grained materials deposited in or adjacent to an ice tunnel by currents, flowing to the south, and (2) A lacustrine facies consisting of fine to very fine sand, silt, and clay, constrained to elevations below the surface of Glacial Lake Winooski (279 m, 915'ft in the south to 295 m, 960 ft in the north).

The esker described in this paper occurs in discontinuous segments extending from 2 km south of Williamstown village to the LePage gravel pit 2 km north of Barre (Optional Stop 5) where it is buried beneath lacustrine sediments. The same ice tunnel and related esker deposits likely extend north up the Kingsbury Branch of the Winooski River through East Montpelier, Calais, and Woodbury and then continued north through Hardwick.

The Stevens Branch Esker is a segmented ridge esker (beaded esker), to use the classification system presented by Warren and Ashley (1994), largely comprised of coarse tunnel fill material and related subaqueous fan deposits. Individual segments of the esker are shown both in map view (Fig. 15A) and in profile (Fig. 15B). Figure 15B shows profile sections of (1) the modern Stevens Branch channel, (2) individual segments of the Stevens Branch esker, and (3) the surface of Glacial Lake Winooski. The cross-section extends from the outlet of Glacial Lakes Williamstown and Winooski approximately 4 km south of Williamstown Village down the Stevens Branch valley to South Barre. The profile of the esker is somewhat altered by gravel excavation, but nevertheless accentuates the discontinuous nature of the esker. Note that in some places the esker ridge rises above the level of Glacial Lake Winooski.

Well-exposed parts of the esker reveal a cross-section consisting of a core of very coarse, poorly sorted, both clast- and matrix-supported, cobble and boulder gravels occurring in lenses elongate in the direction of flow, i.e., parallel to the esker ridge (Figs. 15B and 15C). Poorly developed, large-scale cross-bedding is visible in good exposures and dips both to the south and north. North-dipping cross-bedding may reflect the upstream growth of antidunes. The core of the esker probably formed after short-period, high-discharge flow events where most clasts were entrained within the flow. Deposition was rapid following either or both rapid erosion of the tunnel walls or rapid lessening of discharge in the tunnel. Several of the gravel pits occurring in the valley reveal water scoured bedrock at the base of the esker indicating that flow in the esker tunnel was sufficient to erode all of the underlying till. It is unclear whether significant volumes of till flowed into the ice tunnel, providing an additional source of sediment, from areas adjacent to the tunnel. Eroding till as well as debris melting out of the basal ice are the major sources of sediment in the esker tunnel and discharged from the tunnel mouth.

The core of the esker is mantled by coarse sand and pebble to cobble gravels that are interbedded with medium to coarse sand (Fig. 15C). The cyclic coarse and fine sediments in this part of the esker reflect oscillations in discharge through the tunnel occasioned by weather patterns (warm vs. cold days, rainstorms) and/or seasonal variations in melting. Bedding dips

away from the core of the esker, but small scale cross-bedding indicates current flow was dominantly to the south. Beds of structureless sand and pebble gravel are common suggesting that debris flows were common down the steep sides of the esker, perhaps as support from the surrounding ice was withdrawn. These materials were probably deposited along the sides of the esker core as the tunnel enlarged. This phase of tunnel enlargement may have occurred when the ice was thin enough that it could no longer flow in towards the tunnel during periods of low water pressure.



Figure 15A: Shoreline of Glacial Lakes Williamstown and Winooski is traced on the Barre West and Barre East 7.5-minute quadrangle maps. Isobases (lines of equal isostatic uplift) are drawn at 20 ft intervals from the outlet (915 ft, see arrow on map) south of Williamstown (based on an isostatic uplift gradient of 0.9 m/km, 4.74 ft/mi to N21.5°W; Larsen, 1987b; Koteff and Larsen, 1989). Note that Glacial Lake Williamstown merged with Glacial Lakes Roxbury and Granville when the ice margin retreated north to Barre forming Glacial Lake Winooski. In the Stevens Branch Valley, the shoreline of both lakes are coincident as they shared the same outlet. Trace of Stevens Branch Esker is shown with a dashed line.

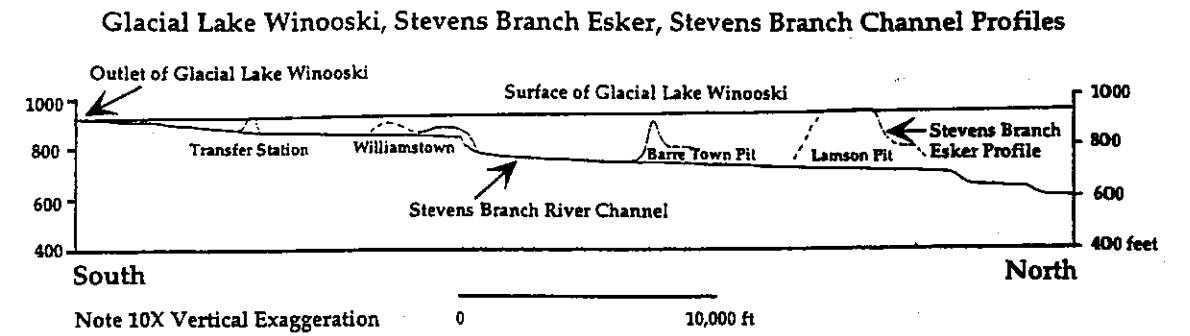


Figure 15B: Profile section of the Stevens Branch Valley (taken from the Barre West 7.5-minute quadrangle map) from the outlet of Glacial Lakes Williamstown and Winooski in the south to South Barre. Section is drawn down the center of the valley with elevations projected into the section where necessary. In addition to the stream profile, the surface of Glacial Lake Winooski (and Williamstown), isostatically rising to the north, and the crest of the Stevens Branch Esker are both shown. Vertical scale has been exaggerated 10X.



Figure 15C: Crest of small esker exposed in lowest excavation of the Lamson Pit, South Barre. View is toward the northeast. Crest of larger esker lies in highest excavations in background. Pebble, cobble gravel is part of the upper tunnel fill facies. This is mantled by medium to coarse sand and pebble, cobble gravel lenses deposited close to or at the mouth of the esker tunnel.

The transition from tunnel-mouth facies to ice-proximal lacustrine facies is evidenced by the rapid decrease in coarse sand and gravel and concomitant increase in medium to fine sand (Fig. 15C). Exposures at the LePage Pit (Optional Stop 5) suggest that pebble/cobble gravels are not transported more than 10's of meters from the tunnel mouth. The esker tunnel also spewed sediment-laden water into more distal parts of the lake. Fine and very fine sand, silt, and clay are distributed throughout the Stevens Branch Valley and are frequently faulted into contact with the underlying ice-contact sediments where both were deposited over buried blocks of ice (Fig. 15D). The medium-fine sand frequently displays ripple-drift cross-lamination, indicating rapid accumulation of sediment. In quieter water, away from the strong currents generated near the ice margin, varved silt and clay were deposited. The clay is much more common in the deeper, northern part of the valley (Wright, 1999b).

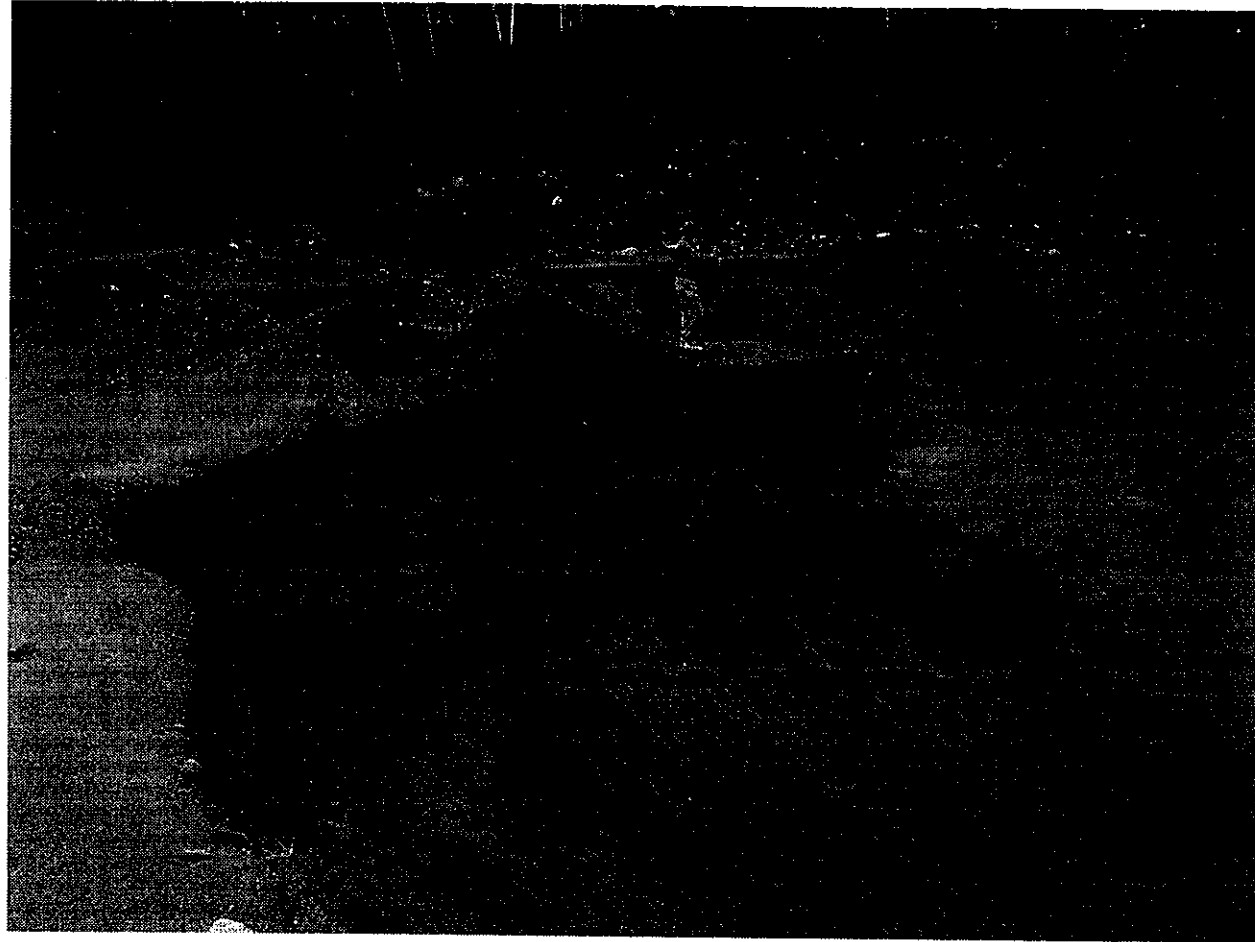


Figure 15D: Coarse sand deposited between the two eskers exposed in the Lamson Pit, South Barre. A dense network of high-angle normal faults indicates that blocks of ice between the two eskers were buried by sand. View looks southeast towards crest of eastern (large) esker.

Summary

The hydraulic gradient in the Laurentide ice sheet generally mimicked the slope of the ice sheet surface (down to the SSE) driving water currents in subglacial tunnels in that direction (Shreve, 1985). The Stevens Branch Esker is probably just one segment of a much longer tunnel system that extended to the north. Abundant cross-bedding in the esker gravels indicates that water currents were flowing southward, up the Stevens Branch valley. The core of the esker, consisting of very coarse, angular cobble and boulder gravel was deposited rapidly. Cyclic high

and low discharge flow is recorded by alternating layers of relatively coarse and fine sediments, although it is likely that only a few of the many cycles experienced in the tunnel are recorded in the sediment record.

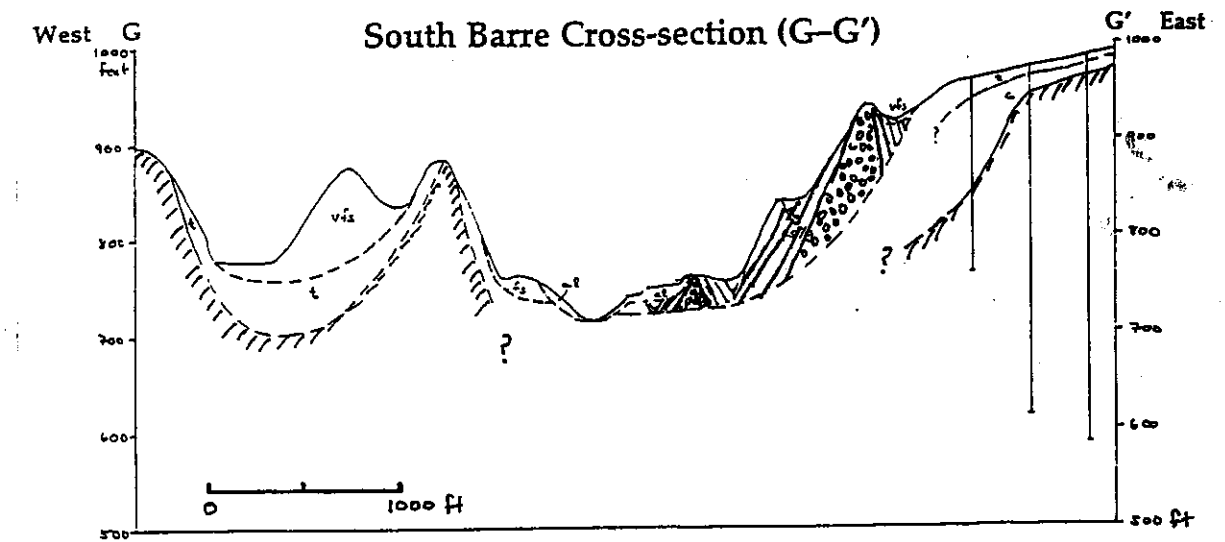


Figure 15E. East-west cross-section of the Stevens Branch Valley, South Barre, extends across the Lamson Pit (see Fig. 15A for location of cross-section). The Stevens Branch Esker is split into two branches, the main branch forming the prominent rise on east side of valley and the smaller branch occurring a short distance to the west. Esker stratigraphy is described in text. Wells are shown as vertical lines on the east side of the section and indicate that till overlies a significant section of lacustrine silt and clay. The till is here interpreted to be a readvance till formed when the ice sheet advanced over lacustrine sediments deposited during its initial retreat down the valley. Bedrock (Devonian Waits River Formation) cores ridges on west side of section. Thick sections of lacustrine fine and very fine sand (fs, vfs) comprise the adjacent hill. Thickness of till (t) and depth to bedrock are poorly constrained away from the wells. Note 5X vertical exaggeration and resulting distortion.

Lamson Gravel Pit, South Barre

Elevation of Glacial Lakes Williamstown and Winooski...~935 ft
 Elevation of crest of Stevens Branch esker.....~940 ft
 Elevation of Stevens Branch (stream)~710 ft

The Lamson pit offers excellent exposures of the Stevens Branch esker over a cumulative vertical section exceeding 60 m (200 feet). The pit exposes two parallel branches of the esker (Fig. 15E). The crest of the main esker lies along the easternmost and highest part of the pit. The crest of the smaller "eskerette" lies immediately east of the entrance to the pit. Medium to coarse sand overlies both eskers and comprises most of the material lying between the two esker crests. This sand is riddled with normal faults indicating that ice blocks were buried between the two eskers (Fig. 15D). The stop descriptions below highlight observations (a) at the crest of the large esker, (b) the area between the two eskers, and (c) the crest of the small esker.

(a) Crest of Large Esker

Recent excavations at the top the pit expose an excellent longitudinal-section of the upper parts of the esker. The core of the esker consists of very poorly sorted/bedded, matrix supported, cobble, boulder gravel. Many of the boulders are still very angular suggesting that they were quickly buried soon after they dropped into the tunnel from the melting ice. Sediments flanking the core consist of pebble gravels with some cobbles and lenses of coarse sand and granules that alternate with fine sand layers. Large-scale cross-bedding dips to the north (upstream or up the hydraulic gradient), suggesting that parts of the core may have accumulated as an anitdune. Flanking beds of coarse sand and pebble/cobble gravel dip to the east on the east side of the esker and to the west on the west side of the esker. Smaller scale cross lamination dip generally south. Small-scale normal faults are common along the west side of the esker.

The crest of the large esker lies at or above the elevation of Glacial Lakes Williamstown and Winooski and thus is not covered by lacustrine sediments (Fig. 2). However, a thin cover of fine sand lies over esker gravels immediately to the east. The fine sand and esker gravels are in contact with till to the east following the 950 foot contour. Recent foundation excavations ~120 m ESE of here exposed lenses of coarse gravel overlying the till that likely originated from the mouth of the nearby esker tunnel. A water well drilled at the same location during the summer of 1998 penetrated a thick section of varved clay beneath the overlying till (Fig. 15E). These are probably lacustrine sediments deposited in a short-lived glacial lake that was overridden by readvancing ice.

(b) Area between Large and Small Esker

This portion of the pit consists entirely of lacustrine sediments that accumulated close to the mouth of the esker. Most of the section consists of thinly bedded fine sand interbedded with very fine sand and silt and lenses of granule gravel. Some beds display load structures and many are partially cemented with calcite. Dense networks of normal faults are common throughout this part of the pit. Most beds are inclined to the southwest which may reflect (1) deposition on the western flank of the large esker, (2) tilting by faulting, or (3) the structure of a small subaqueous delta that developed in front of the esker tunnel.

(c) Small Esker

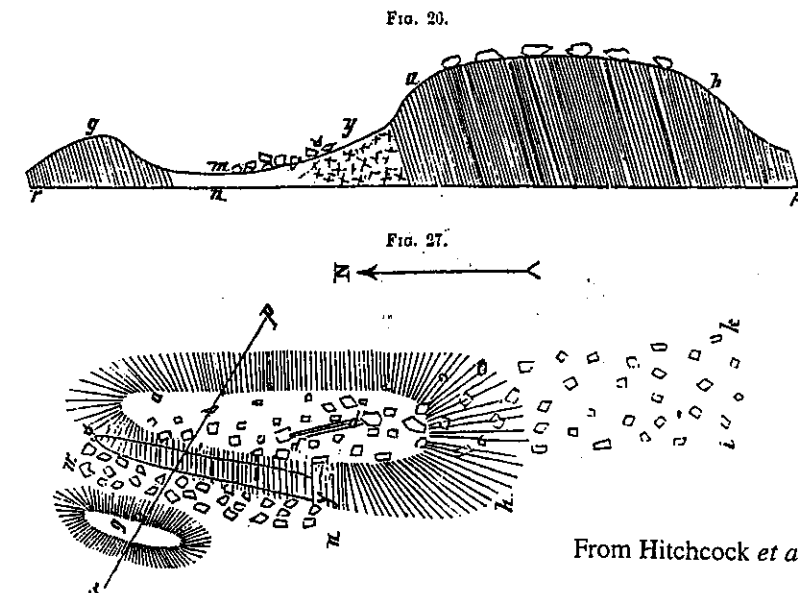
The small esker is currently exposed in two faces, one, immediately east of the pit entrance and the second in the deepest pit excavation a short distance south of the pit entrance (Fig. 15C). The core of the small esker consists of cobble/boulder gravel overlain by poorly bedded pebble and cobble gravel that arches over the boulder gravel core. The esker gravels are covered by lacustrine sediments draped over the underlying crest of the esker (Fig. 15C). These consist of alternating cm-scale layers of silt and coarse sand with occasional lenses of pebble gravel, that were probably deposited relatively close to the mouth of the esker tunnel. Deformation due to slumping and contracting tunnel walls is common.

- Proceed north on VT Rt. 14 53.6 traffic light at junction with VT Rt. 63
 56.7 famous Scottish restaurant with yellow arc-like structures
 54.6 Quarry Hill Road, ROAD TO BANQUET
 54.9 Traffic light, bear right on Hill St, which bends to the right at 55.05
 55.3 Turn right (east) on U.S. Rt. 302, continue to the east out of Barre
 57.6 Good view of Jail Branch Section, till over lacustrine sequence (Larsen, 1972, Stop 1)
 59.1 East Barre flood control dam 59.8 Turn left (north) on Reservoir Road
 62.1 Thurman Dix Reservoir on the left 64.1 Cutler Corner Road on left, drainage divide
 65.3 Gore/Maxfield Roads intersection 66.1 New section exposed by slump at log house
 66.7 Park on left for Stop 7

TRAINS OF BOWLERS.

Fifteen years ago we gave an account in the American Journal of Science, of some remarkable trains of angular bowlders, strewed almost in a strait line over the Taconic ranges of mountains southeasterly, some ten or fifteen miles, especially in the town of Richmond in Massachusetts. They were remarkable for not being rounded at all, for being strewed along as if by art in nearly straight lines, and with well defined borders, and also for lying above the common rounded bowlders. Subsequently they excited much interest, and were described anew by Prof. Henry D. Rogers and Sir Charles Lyell. In 1846 Prof. Adams discovered and described a similar case on land of Mr. Butts, in Huntington, Vermont, pointed out to him by Mr. Bunyan Bradley. One of our number has visited the place, but we prefer to re-produce Prof. Adams' figures (Figs. 26 and 27) and description, with a few slight corrections.

a h is a hill of talcose slate 150 feet high, and *g* a small hill of the same rock. *s y* is a narrow strip of syenite or perhaps hornblende rock, interstratified with the slate, numerous blocks of which have fallen into the valley below, *m n*, and are strewed over the highest hill and beyond it, as represented in figure 27, to *i k*, for a mile and a half. On the top of the hill are striae, *d d*, running N. 15° W. as shown on the figure, and the line *p r* shows where the section (Fig. 26) crosses the two hills. That section will show the dip and position of the rocks and the bowlders.



From Hitchcock *et al.* (1861, p. 64).

STOP 7A: GREAT BROOK, LODGEMENT TILL OVER VARVES (Springston)

A landslide scar on the west side of Great Brook exposes lodgement till overlying varved lacustrine material ranging from fine sand to clay. Due to slumping of the deposits the base of the lacustrine material is not exposed, but nearby outcrops of lodgement till at stream level strongly suggest that this underlies the lake deposits. The brook here is at an elevation of 1120 ft ASL (341 meters), far above the elevation of glacial Lake Winooski, which only extended up this valley to about 960 feet (293 meters). Thus the lacustrine material must have formed in a high-level proglacial lake within the northward draining Great Brook valley. As no absolute ages have been assigned to any of the units exposed at this stop, it is unclear whether the lake formed during advance of the late Wisconsinan glacier or during the Middlesex Readvance described by Larsen (2001). This readvance will be discussed in greater detail at Stops 8 and 9.

A schematic stratigraphic section is shown in Figure 16. The upper lodgement till is about 14 feet (4.3 meters) thick. The lacustrine material is approximately 19.5 feet (5.9 meters) thick, with the upper 4.5 feet (1.4 meters) deformed. The remainder of the exposed lacustrine deposit is undeformed and ranges from fine sand to silty clay and clay. A sequence of 9 thick varves was measured, ranging from 0.15 feet (45 mm) to 1.88 feet (.57 meters) in thickness, with the final 3.60 feet (1.1 meters) in the undeformed section appearing to make up a single varve. The internal stratigraphy of each individual varve consists of "summer" layers that contain numerous laminae that fine upward from fine sand to silty clay, and each of these summer layers is capped by a thin winter layer of nearly silt-free clay. Landslide deposits cover the lower 12 feet of the main section but a small exposure 75 feet upstream of the main slide scar shows identical varved lacustrine material resting on lodgement till. However, the situation is complicated by the fact that the varves also are in abrupt contact with lodgement till several feet to the left and with loose medium sand on the right, probably a result of ice-contact deformation. Several exposures of lodgement till are seen in the brook about 400 to 500 feet downstream from this site. This is the locality for the lower lodgement till described in the next paragraph.

Comparison of the Two Tills

The lower till at this site is the typical lodgement till found throughout this part of Vermont. In the unweathered state it is dense to very dense, with a silt-sand matrix and subrounded to rounded pebbles, cobbles, and boulders that are commonly striated. At some localities it contains a few discontinuous lenses of sand. The color is typically dark gray (Munsell N4/0). The fresh till is calcareous and often has a sub-horizontal fissility. Weathered exposures have lost much of the carbonate, are of a looser consistency, and typically have an olive color (Munsell 5Y4/2 to 5Y4/4). The upper till at this site is looser and more olive than the typical lodgement till in this area but this difference may simply be due to weathering. The next landslide that occurs at this site may well expose fresher material.

As shown in Table 1, the clast lithologies are similar in the two tills. Samples GB-44U and GB-44L are the upper and lower tills at this stop, respectively. GB-178 is a typical lodgement till at a site located 0.25 mile (0.42 kilometer) north of Stop 7A. Samples GB-2A and 3A are from Stop 7B, described below. The most common lithologies observed are calcareous granulite, rusty schist and phyllite, and massive gray quartzite (all to be found in the rocks between here and the Taconian Line or RMC). Vein quartz clasts are present in all the samples and may be of local

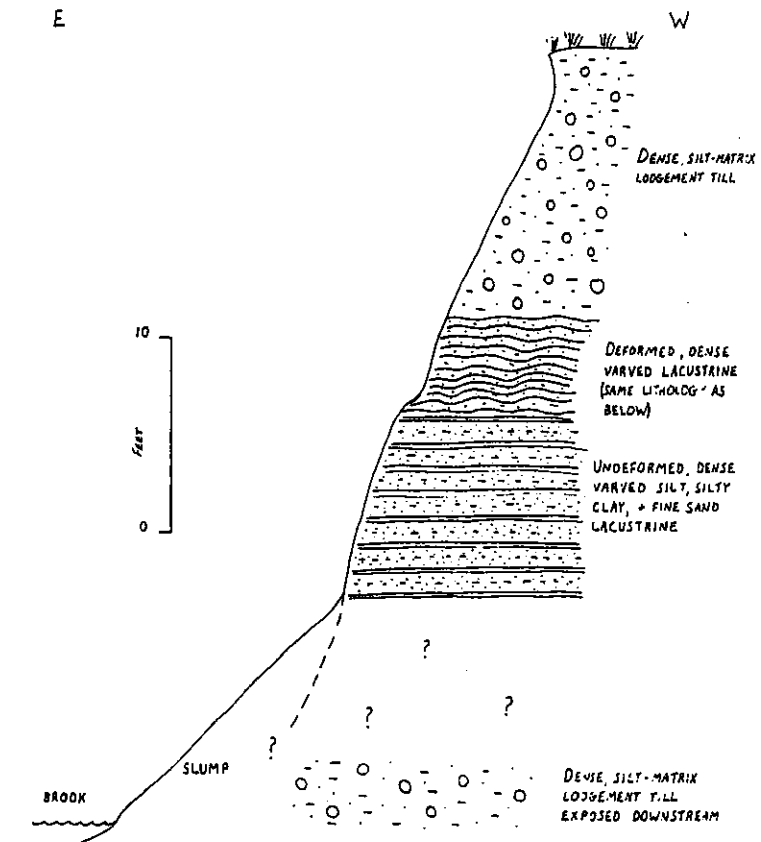


Figure 16. Stratigraphic Section at Stop 7.

origin. The granite clasts may be partly from local granite exposures (small sills or dikes) while others may be from granite bodies 6 to 10 miles northwest of the watershed near the village of Adamant in the town of Calais. The amphibolite clasts are mostly similar in appearance to the metamorphosed mafic rocks which occur in the local Waits River Formation. The nearest exposures of greenstones are located approximately 13 miles to the northwest and the nearest serpentine body is located approximately 17 miles to the northwest.

Most of the clasts are of local origin. As there is abundant granite in the hills to the northeast of the study area, the low granite content in the till provides a strong argument against ice movement from the northeast.

Ice Movement Directions

Glacial striations were observed at two sites in the study area. At Station GB-178 striations on an outcrop in Great Brook have an orientation of 160° . At a ledge in the brook located 0.15 mile (0.25) kilometer south-southeast of the intersection of Brook Road and East Hill Road striations have an orientation of 165° . Striations located to the south of the study area in the Barre East quadrangle (on ledges adjacent to Nelson Brook, 1560 feet elevation) have orientations of 180° to 177° .

Table 1. Lithologies of till clasts at five localities in the Great Brook watershed. Expressed as percentages of pebbles counted.

Lithologies	Localities				
	GB-2A	GB-3A	GB-178	GB-44U	GB-44L
Calcareous granulite	78	50	42	44	47
Schist and phyllite	4	32	26	28	34
Quartzite	2	8	10	10	4
Foliated quartzite			4		
Red quartzite					1
Granite	4	4	6	2	
Vein quartz	8	4	4	8	5
Amphibolite	4	6	4	2	
Greenstone				8	8
Serpentine		2	4		1

N = 50 for each locality except GB-44L = 100

The direction of ice movement based on glacial erratics in the Great Brook watershed is consistent with the striations. A large boulder of serpentine was discovered in the brook (Mile 66.9 in Road Log) and three serpentine boulders were seen in the brook at a site 0.45 mile (0.75 kilometer) NW of the intersection of Brook Road and East Hill Road. An erratic cobble of the Craftsbury orbicular granodiorite (Doll and others, 1961; Konig and Dennis, 1964) was also discovered in the brook. To move this cobble from the source area in Craftsbury into the watershed at or upstream of this station requires a movement line between 176° and 186° , which is in reasonably good agreement with the striations described in the previous paragraph.

Till fabrics were measured in the upper and lower tills at Stop 7A and in the upper till and lower waterlain till at Stop 7B described below and are shown in Figure 17. The upper till at Stop 7A shows a maximum at about 165° but considerable scatter in other orientations. The lower till has a prominent maximum at about 195° and less overall scatter. The valley at these sites has an overall trend of about 170° . Fabrics at Stop 7B are discussed below.

Proceed north (downstream) 67.5 15 foot serpentine boulder on left
68.0 Site of house washed away in 1984 flood 69.4 Slow, park on left for Stop 7B

STOP 7B: GREAT BROOK LODGEMENT TILL OVER WATERLAIN TILL AND FINE LACUSTRINE SEDIMENTS (Springston, Optional)

At this site two large landslide scars on the west side of Great Brook expose fine grained lacustrine silty clay, silt, and silty fine sand that are interbedded with diamict that has the consistency, texture, color, clast shape, clast lithologies, of the lodgement till. Based on the numerous beds of these two materials, the diamicts are interpreted to be rain-out deposits formed from the melting of icebergs in a proglacial lake. These materials are, in turn, overlain by a layer of lodgement till that is to all appearances physically identical to the diamict underneath and the tills examined at Stop 7A. Starting at the level of the brook, the section consists of 38 feet of the waterlain till and lacustrine silt and silty clay. These are overlain by 14 feet of lodgement till that is increasingly weathered in the top several feet of the section.

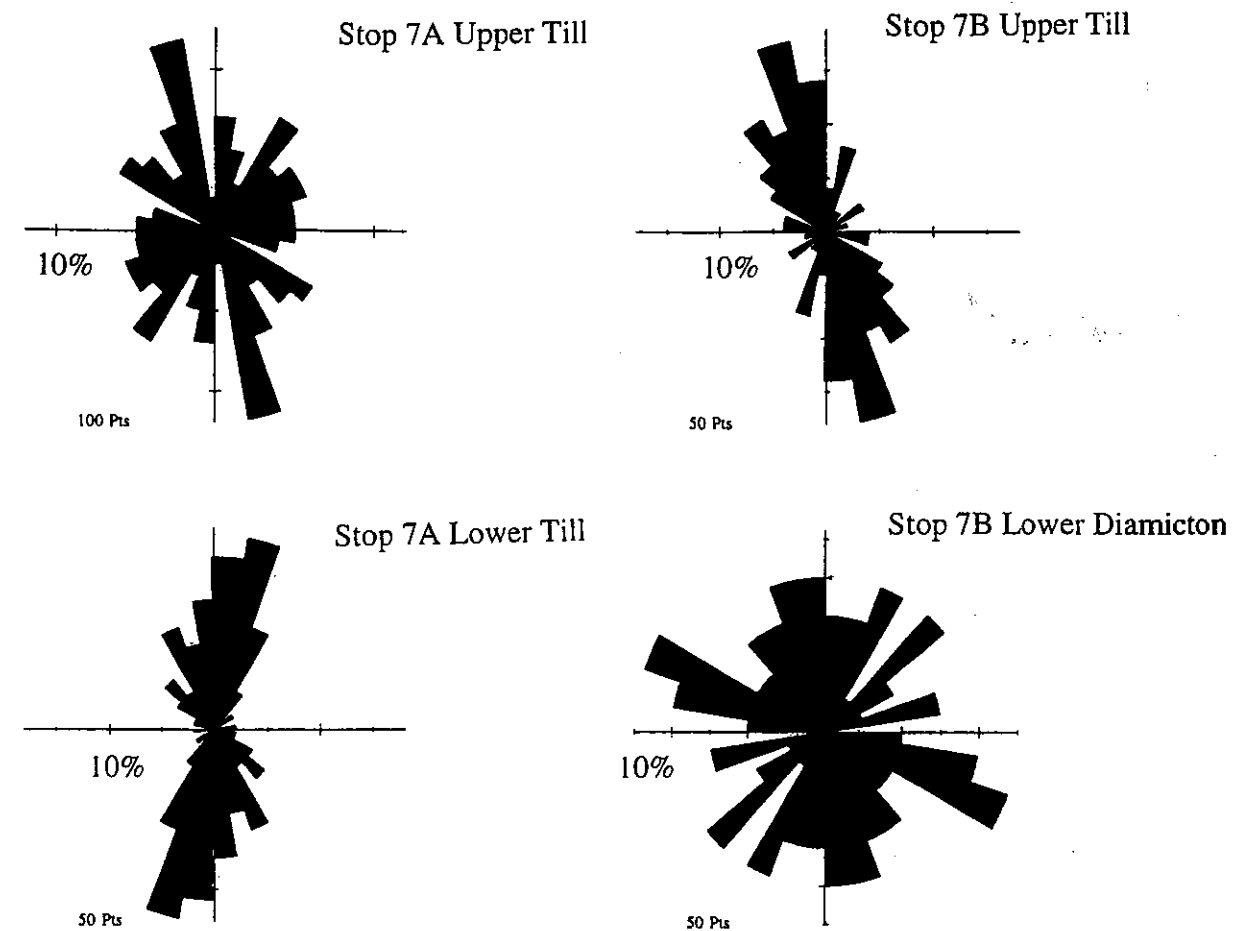


Figure 17. Till fabric diagrams for Stops 7A and 7B. North is up.

Clast lithologies in the upper till and the waterlain till are shown in Table 1 as samples GB-3A and GB-2A, respectively. As at Stop 7A, most are of local origin. Again, note the absence of abundant granite, indicating that the ice did not come from the northeast.

Till fabrics are shown in Figure 17. The upper till at this site has a strong maximum at about 160° . This is consistent with the glacial striations and erratics described above. The waterlain till at the base of the section has a very weak preferred orientation, as would be expected of material that accumulated as debris rained down through the lake waters from a floating ice tongue.

The brook at this stop is at approximately 768 ft ASL (234 meters); well within the projected level of glacial Lake Winooski at approximately 960 feet (293 meters). However, the presence of the till at the top of the section indicates that the lacustrine part was deposited prior to the final retreat of the ice from the Plainfield area. The question is which lake was this material deposited in? There are two possibilities: a proglacial lake that predated the Late Wisconsinan glaciation and then would have been overridden by the main ice sheet or a lake that formed in the Great Brook valley or the valley of the upper Winooski River during the Middlesex Readvance of Larsen (2001).

- Continue north, downstream
- 69.7 Turn right (north) at stop sign; At 69.85 turn left, cross bridge over Winooski River
- 69.9 Stop sign, turn left on U.S. Rt. 2, proceed west through village of Plainfield
- 73.25 Stop sign, join Rt. 14, turn left in village of East Montpelier
- 73.5 Vt. Rt. 14 leaves on left, continue west on U.S. Rt. 2
- 74.2 turn right (west) on Town Hill Road; enjoy Vermont scenery, view of Camels Hump
- 77.7 Stop sign, turn left on Main St; New England Culinary Institute on the right
- 78.5 Roundabout, YIELD to traffic on roundabout, bear right, join VT Rt 12
- 78.6 Stop sign, turn right (north) on Elm St (= VT Rt. 12)
- 79.7 Slopes west of Montpelier swimming pool are underlain by varved silt and clay of Lake Winooski. In 1999, new exposures at Woodbury College had bedrock, till, and a thinning-upward sequence of fine sand, silt and clay
- 80.2 North Branch Nature Center, northern outpost of VINS (Vermont Institute of Natural Science, erosional topography to the east is underlain by varves capped by terrace/fan deposits
- 80.5 Striations on polished Moretown "pinstripe" trend S22°E
- 82.25 Wrightsville Dam on right, built in 1933-35 by Civilian Conservation Corps following 1927 flood
- 82.32 WITH CAUTION TURN LEFT ACROSS TRAFFIC to parking space on west side of Rt. 12. PLEASE USE CAUTION IN CROSSING ROUTE 12 ON FOOT. Proceed to northeast corner of field opposite parking area and descend wet, slippery slope.

STOP 8: UNDISTURBED VARVES OF LAKE WINOOSKI (Larsen and Wright)

This is Stop 1 in the 1999 NEIGC Field Trip Guidebook (Larsen, 1999b, p. C1-11). The base of the section is about 60 ft above the reservoir, which is about 20 ft deep. At the base of the section fine sand rests on till (Figs. 18 and 19). The absence of coarse gravel at the till-sand contact indicates that there was no subglacial stream near this site when the ice margin retreated through the area. In sharp contact above the sand are 14.9 ft of thin sticky varves that formed in a quiet Lake Winooski. The varves are soft (i.e. not consolidated), plastic and easy to excavate. The bedding is flat and dips at a low angle toward the center of the valley. Stephen Wright and his students measured 183 varves at this site and their varve curve matches a portion of Antevs' (1928) New England varve curve for the Winooski valley (Larsen et al, 2001). Resting above the varves is pebble gravel with a well-developed soil profile covered by artificial fill. The pebble gravel was mapped as "terrace/fan deposits" and was formed by Culver Brook when Lake Winooski drained.

Return to cars and proceed north, WATCH FOR SPEEDING CARS IN RE-CROSSING AND CROSSING ROUTE 12 IN JUST 0.05 OF A MILE.

- 82.4 Turn left on Culver Hill Road
- 82.8 Turn left and park on Warren Road; walk south 220 ft, drop down over bank, follow trail southeast to Stop 9

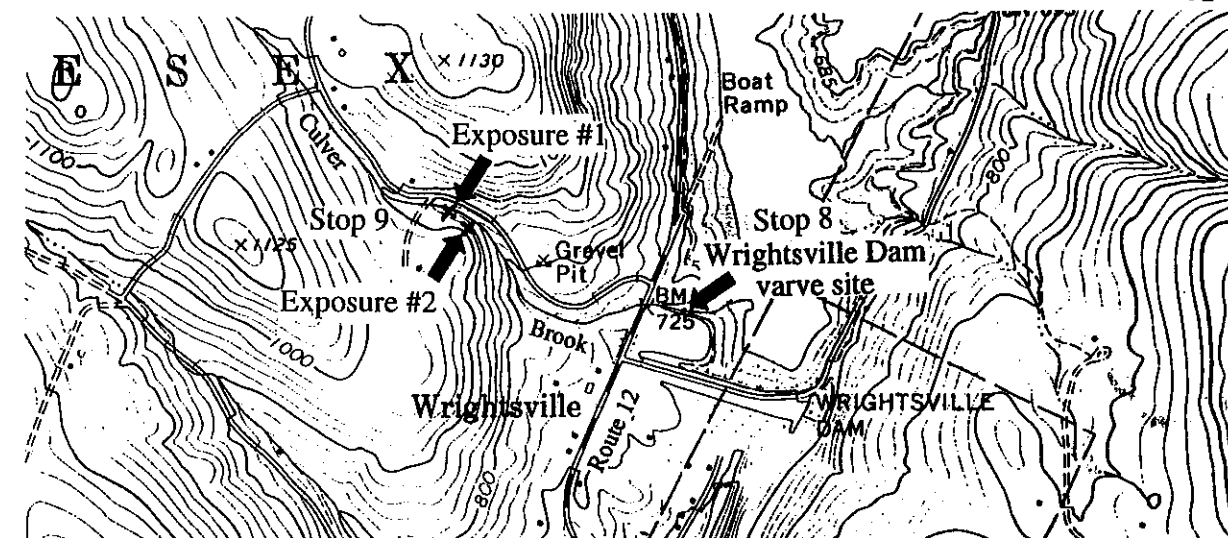


Figure 18. Topographic map showing the location of Stops 8 and 9.

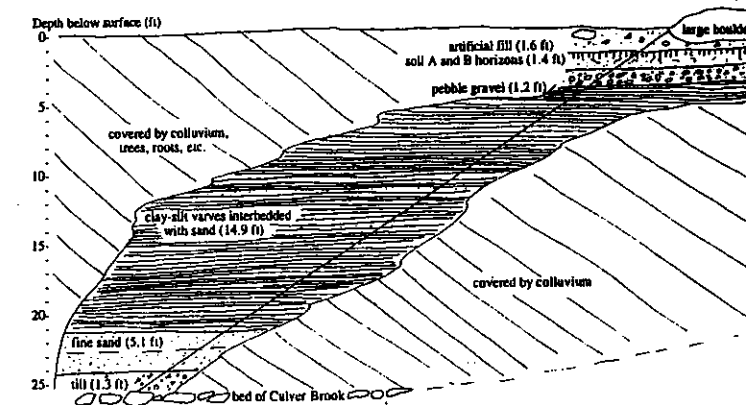


Figure 19. Stratigraphic section looking south at undisturbed varves of Lake Winooski at Stop 8. Section prepared by the Glacial Geology class at Norwich University on 9/9/99.

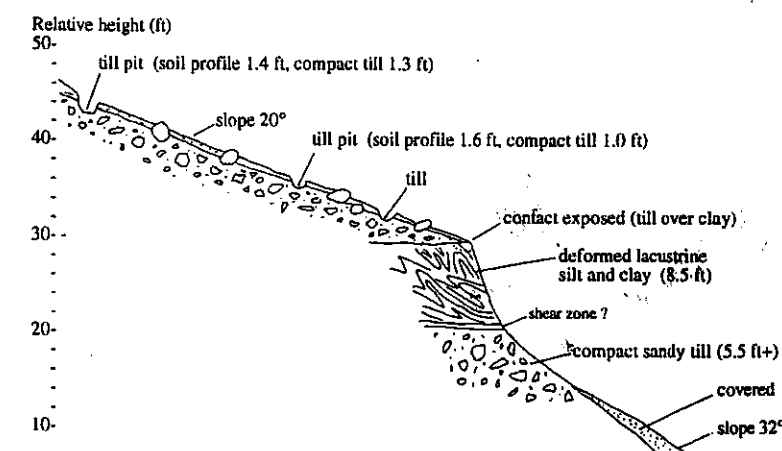


Figure 20. Stratigraphic section looking northwest at Stop 9, Exposure #2.

STOP 9: CULVER HILL ROAD, TILL-VARVES-TILL SECTION (Larsen)

This is Stop 5 in the 1999 NEIGC Field Trip Guidebook (Larsen, 1999b, p. C1-13). The first of two exposures is adjacent to Culver Brook and consists of two tills. The upper till is gravelly, loose to compact, brown to gray, overlies a sharp contact on the lower till, and contains slabs or clasts of both lower till and lacustrine sediments. The lower till is very compact and has a dark greenish-gray to bluish-gray color when moist, similar to rocks in the Moretown Formation. The upper 3.3 ft of the lower till is light to dark yellowish-brown and was interpreted by the senior author in 1999 to be a weathered zone on an Early-Wisconsinan till. However, organic material in a slab of fine to very fine sand in the upper till has an age of $11,900 \pm 50$ ^{14}C years BP (Geochron No. GX-26457-AMS), which is the same age assigned by Ridge et al (1999) for the Bethlehem readvance at Comerford dam. This is the type locality for the Middlesex readvance (Larsen, 2001b). To reach the second exposure, climb up the slope at the right end of the first exposure and follow trail on the contour about 200 ft to a landslide scar. Watch your footing, the slope is slippery.

Exposed in the landslide scar is a package of blue-gray, very compact and highly deformed clay-silt varves resting on yellowish-brown, compact, sandy till that is similar to the lower till at the first exposure. Between the deformed varves and the till is a 4 in layer of light yellowish-brown, fine to very fine sand and silt, which appears to be a shear zone on which the overlying package of varves moved. Three shallow pits were dug on the slope above the landslide scar and yellowish-brown, compact till was exposed in each. This site has two tills separated by deformed lake sediments (Fig. 20). The lower till was deposited during the Late Wisconsinan ice advance and retreat. When the ice margin retreated in this area, Lake Winooski varves were deposited. Some of those lake sediments were glaciotectonically transported to this site during a readvance of the ice margin about $11,900^{14}\text{C}$ years BP at the same time that ice readvanced at Comerford dam.

Return to cars, proceed east on Culver Hill Road

83.3 Turn right, follow VT Rt 12 to Econo Lodge. To avoid downtown Montpelier, follow Elm St south to State St, turn right, proceed 0.1 mi west, turn left on Taylor St, cross steel girder bridge over Winooski River, turn left (east) on U.S. Rt 2 at stop light, proceed 0.2 mi to stop light, turn right on VT Rt 12, proceed 0.4 to Econo Lodge.

ROAD LOG - DAY 2

0.0 mi, 8:00 AM, depart from south parking lot at Econo Lodge, proceed west, follow Day 1 road log to Interstate, I-89, turn west; greenish-gray outcrops for next 4 miles are quartzite, phyllite and greenstone of the Ordovician Moretown Formation

12.2 Turn right, leave I-89 at Exit 10

12.5 Stop sign, turn left on VT Rt. 100

13.0 Turn right (west) on U.S. Rt. 2

14.4 Turn right (north) under I-89 on Little River Road

17.3 Waterbury Dam, reservoir is low while dam is under repair

17.9 Entrance Station, Little River State Park 18.6 End of road, park cars, Stop 10

STOP 10: Waterbury Reservoir Section (SFW)*Introduction*

The Waterbury Reservoir dams the Little River approximately 3 km north of its confluence with the Winooski River (Fig. 21). The reservoir was constructed between 1935–1938, in response to both the depression and the significant damage sustained within the Winooski valley during the 1927 flood, the largest flood of record in northern Vermont. The reservoir flooded the community of Little River near the confluence of Stevenson Brook and Bryant Brook with the Little River (also called the Waterbury River on older maps; Fig. 22). During the 1920's Ernst Antevs measured two varved silt/clay sections exposed by landslides within the bounds of the present day reservoir (Sections 151 and 152, Fig. 22; Antevs, 1928). Antevs used these and 3 other sections farther east in the Winooski river valley to construct his normalized Winooski Valley varve curve (20 VT, Antevs, 1928). This field stop describes a well-exposed section of glacio-lacustrine sediments located close by Antevs' Section 152. The water level behind the reservoir has been lowered between 40 and 70 feet below the normal pool elevation of 590 feet during the last few years to facilitate repairs to the dam. The reduced water levels have exposed the lower part of the glacial lake section as well as modern reservoir sediments. Stop 10 lies within the Little River State Park. Access to the section is from the boat launching area.

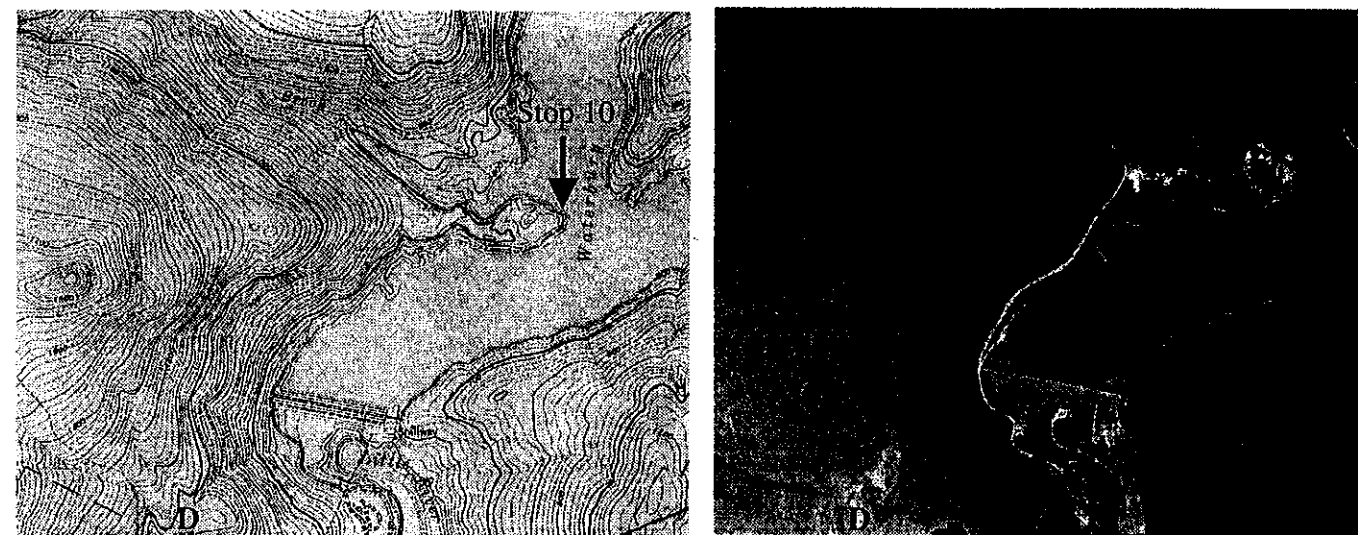


Figure 21: Portion of the Bolton Mountain 7.5-minute Quadrangle (1948) and aerial photograph of the Waterbury Reservoir (1962; VT-62-H 29-115) showing location of the stratigraphic section described in the text (Stop 10). Note that access roads to the campground are not shown on the topographic map, but appear in the photo. Dark line at 1,020 ft outlines the shoreline of Glacial Lake Winooski. Dark line at 690 ft outlines the shoreline of Glacial Lake Mansfield. "D" on photo and map is a Lake Winooski Delta. Reservoir dam is ~700 m long.

Location and Geologic Setting

The reservoir area is underlain by the Hazens Notch Formation, a late Precambrian to Cambrian NNE-SSW-trending belt of metasedimentary and metavolcanic rocks consisting predominantly of rusty weathering (pyrite), graphitic schists, quartzites, and gneisses (Stanley et al., 1999). While not exposed at this field stop, bedrock knobs occur within 100 m of the stop and show considerable local relief. The broad north-south valley drained by the Little River (extending north towards Stowe and Morrisville) is controlled by the preferential weathering of the Stowe and Ottauquechee Formations, whereas the WNW-ESE trending Winooski River valley and the eastern arm of the Waterbury Reservoir cut across formation boundaries and are structurally controlled.



Figure 22: Pre-reservoir topographic and cultural setting of the Little River valley from the Camels Hump (1915) and Montpelier (1919) 15-minute Quadrangle maps. The Waterbury Reservoir section described here (Stop 10) lies just above the "0" in "520," immediately north of "Little River School." Antevs' (1928) section 152 is located just south of Stop 10 (A152) and Antevs' section 151 is located along Bryant Brook, now the eastern end of the reservoir (A-151).

The broad glacial history of the Winooski River valley has been described earlier in this guide. Here, as in most of the Winooski River drainage basin, the retreating ice front dammed a series of glacial lakes: Glacial Lakes Winooski and Glacial Lake Mansfield. Locally Glacial Lake Winooski was the first glacial lake to exist. Differential isostatic uplift to the NNW has tilted the lake surface from 915 ft at its outlet to a local elevation of ~1020 ft near the Waterbury Reservoir. This lake grew rapidly to the north where it occupied the Lamoille valley which was also dammed by glacial ice (Wright et al., 2001). Glacial Lake Winooski drained when ice retreat uncovered a lower outlet through the Huntington River Valley (Larsen, 1987b). Uncovering this outlet lowered lake levels by ~330 ft (100 m) creating a new lake called Glacial Lake Mansfield. This new lake was at a local elevation of 690 ft (Fig. 21) and existed until the ice sheet retreated out of the Winooski River valley and the valley became an arm of Glacial Lake Vermont. The Coveville stage of Lake Vermont was at a local elevation of ~550 ft, below the elevation of Stop 10. The section exposed along the Waterbury Reservoir reveals some of the history of Glacial Lakes Winooski and Mansfield in this part of the valley.

Description/Interpretation of the Section

The section we will visit is located on the easternmost end of a peninsula protruding from the western shore of the Reservoir towards the center of the Little River valley (Fig. 21). Prior to the formation of the Reservoir the peninsula was bordered by Stevenson Brook to the north and the Little River to the east and south (Fig. 22). Recent landslides have revealed several sections along the peninsula; it is the northernmost of these that is described here (Fig. 23).

Figure 23: Landslide scar exposing section along the easternmost end of peninsula extending into the Waterbury Reservoir. Lowermost part of section (here covered by landslide debris) consists of ice-proximal fine sand with silt interbeds. This is overlain by a section of varved silt/clay (Fig. 25, above fallen tree). There is an abrupt transition halfway up the section from varved silt/clay to fine to coarse sand with silt interbeds. The uppermost part of the section consists of coarse alluvium.

The section consists of four distinct packages of sediments (Figs. 23 and 24). (1) The lowest of these consists of thickly bedded (meter-scale) fine and very fine sand interbedded with much thinner (<10 cm thick) beds of silt. This fine sand is at least 20 m thick extending from the normal reservoir level (590 ft asl) to and below the lowered reservoir elevation (550 to 520 ft asl; Fig. 24). Nearby sections along the southern margin of the peninsula show that the fine sand is deposited directly on water-worn bedrock, i.e. no till or coarse ice-contact sediments separates bedrock from the overlying fine sand in this area. At the lowered reservoir level (~550 ft asl) the shoreline of the reservoir is bordered by either bedrock or fine sand. Powerboat wake and wind-generated waves rapidly resuspend this sand near shore. (2) A low-angle normal fault and slumped silt/clay separate the fine sand from approximately 6 m of varved silt and clay. (3) There is an abrupt change from the varved silt and clay to a coarsening upwards sequence of sand with silt interbeds that makes up most of the upper half of the exposed section. (4) The uppermost part of the section consists of lenses of alluvium (coarse sand, pebbles, and cobbles) that unconformably overlies the medium to coarse sand section below. The entire section is cut by normal faults.

The lowermost fine to very fine sand portion of the section was deposited in an ice-proximal environment (within several km of the mouth of a subglacial tunnel). During ice retreat the most long-lived ice tongue slowly retreated to the west, down the Winooski River Valley and pumped extensive volumes of sediment into Glacial Lake Winooski (Larsen, 1987b). An ice tongue also retreated to the northwest, up the Millet Brook valley (approximately 8 km to the north of the Reservoir) depositing an esker system that was also coeval with Glacial Lake Winooski (Wright et al., 1997). This ice tunnel was also a source for sediments entering Glacial Lake Winooski in the vicinity of the Reservoir.

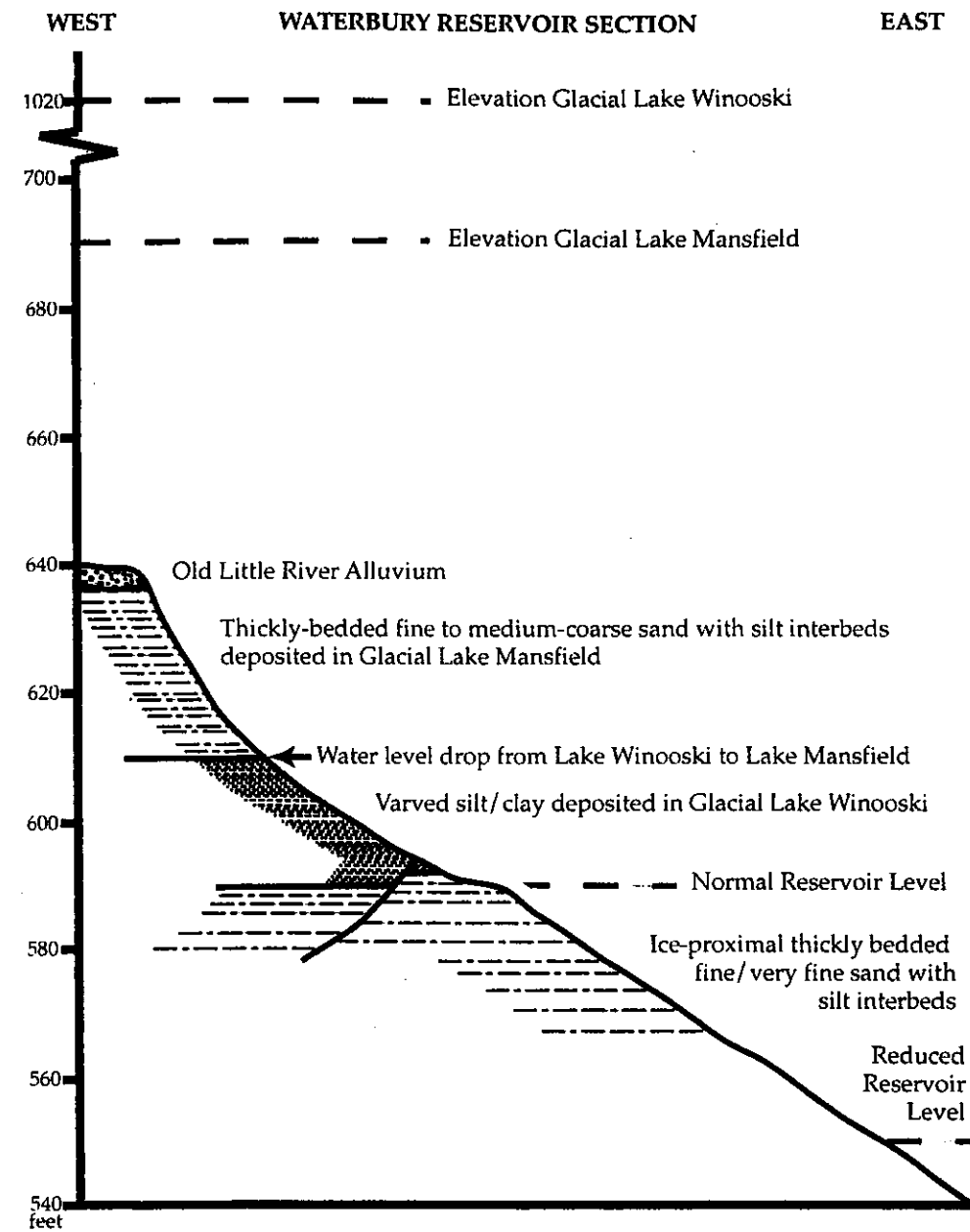


Figure 24: Generalized cross-section of the stratigraphic section described in the text (see Fig. 21 for location). Slopes are approximate. Note scale break for "Elevation of Glacial Lake Winooski."

The varved silt/clay portion of the section was measured during the fall of 2002 and is presented in Figure 25. There is an abrupt change between the thickly bedded fine sand/silt section below and the thin (2-3 cm thick) silt/clay varves at the base of the measured section (Figs. 24 and 25). Extensive faults and slumped bedding indicate that part of the section is missing, presumably slumped into the deeper parts of the lake. Stratigraphically, the missing section represents the transition between the ice-proximal section below and the ice-distal section above. Antevs (1928) section 151 along Bryant Brook, ~2.7 km west of this section, contains 46 additional couplets at the base of the section suggesting that a significant portion of the section is missing.

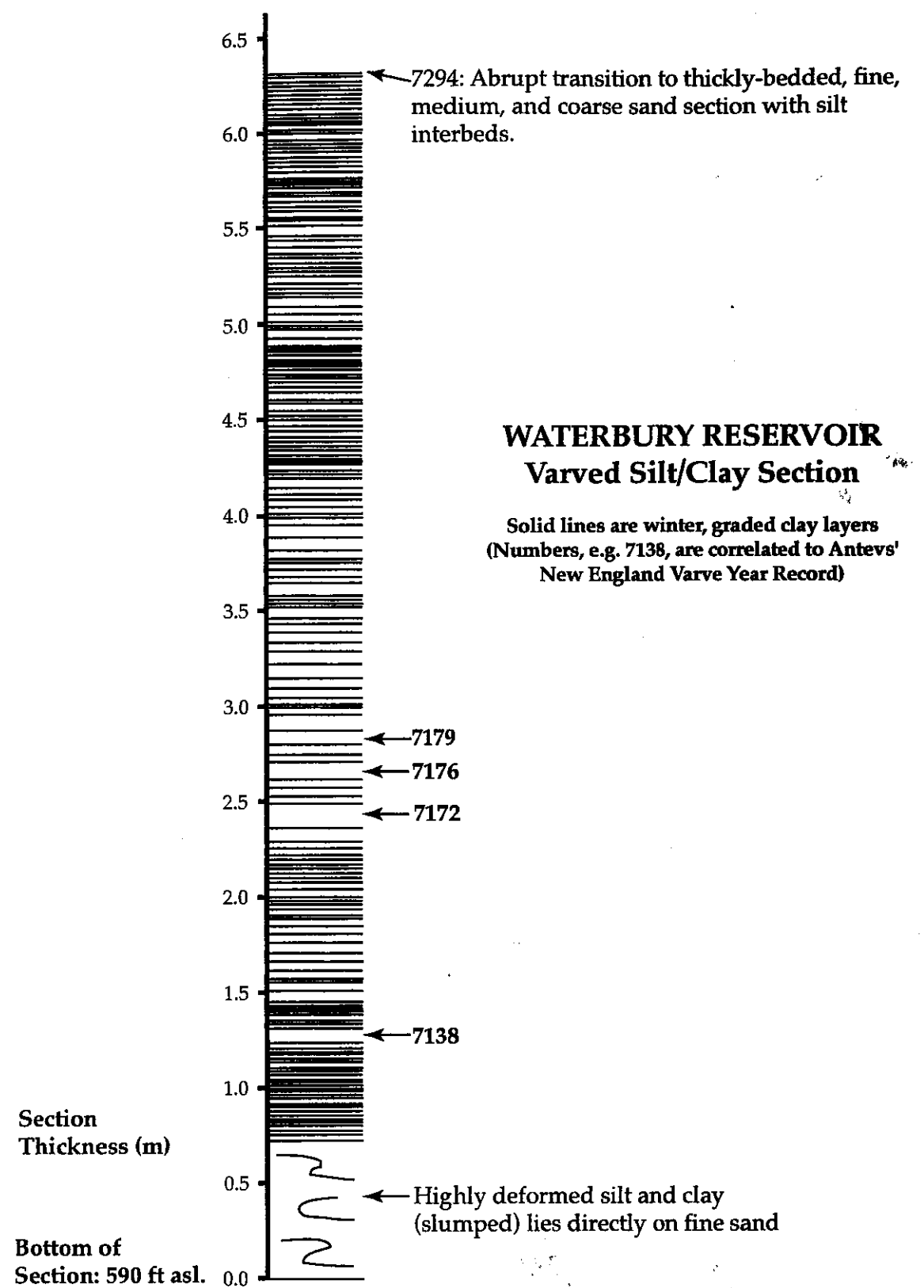


Figure 25: Measured stratigraphic section of silt/clay varves from exposure along the western shore of the Waterbury Reservoir (see Fig. 21 for location). Base of section begins at the contact between fine sand below and slumped clay/silt above which lies close to the normal reservoir elevation of 590 ft asl. Top of measured section ends with the first influx of thick medium sand beds, corresponding to the drop from Glacial Lake Winooski to Glacial Lake Mansfield. Numbers, e.g. 7138, refer to a specific New England Varve Year (Antevs, 1922, 1928) modified by Ridge et al. (1999).

Most of the winter clay layers are maroon colored, are graded and generally easy to distinguish from the summer silt layers. The clay layers are also frequently jointed. Low-amplitude ripples occur at the base of some couplets (first spring influx of sediment into the lake) and all indicate currents flowing from north to south. Rare plant fossils have been recently found within some of the silt layers (Fig. 26) indicating that the local area was being revegetated during the lifespan of Glacial Lake Winooski. Worm tracks are also common as are concretions. Individual layers within the varved silt/clay section show extensive syn-sedimentary soft-sediment deformation.



Figure 26: Photograph depicts a single bedding plane containing an unidentified sprouting seed preserved in silt layer approximately 2 m from base of section.

The measured silt/clay section contains 181 couplets. These have been correlated with Antevs' (1928) type Winooski valley section (Fig. 27). A distinctive pattern of thick varves, beginning with varve number 7138, has facilitated the correlation (Figs. 25 and 27). Based on the more complete section measured by Antevs (1928) less than 3 km to the east (Section 151), Glacial Lake Winooski lasted for a minimum of 227 years in this area. To this span of time must be added the time necessary to accumulate the ice-proximal fine to very fine sand that underlies the silt/clay couplets.

The water level of Glacial Lake Winooski fell approximately 100 m (330 ft) when the ice retreated far enough west-north-west, down the Winooski River valley, to expose the Hollow Brook outlet in South Hinesburg. The newly formed, lower-elevation lake that occupied the

Correlation of Waterbury Reservoir Section to Antevs' Winooski Valley Section

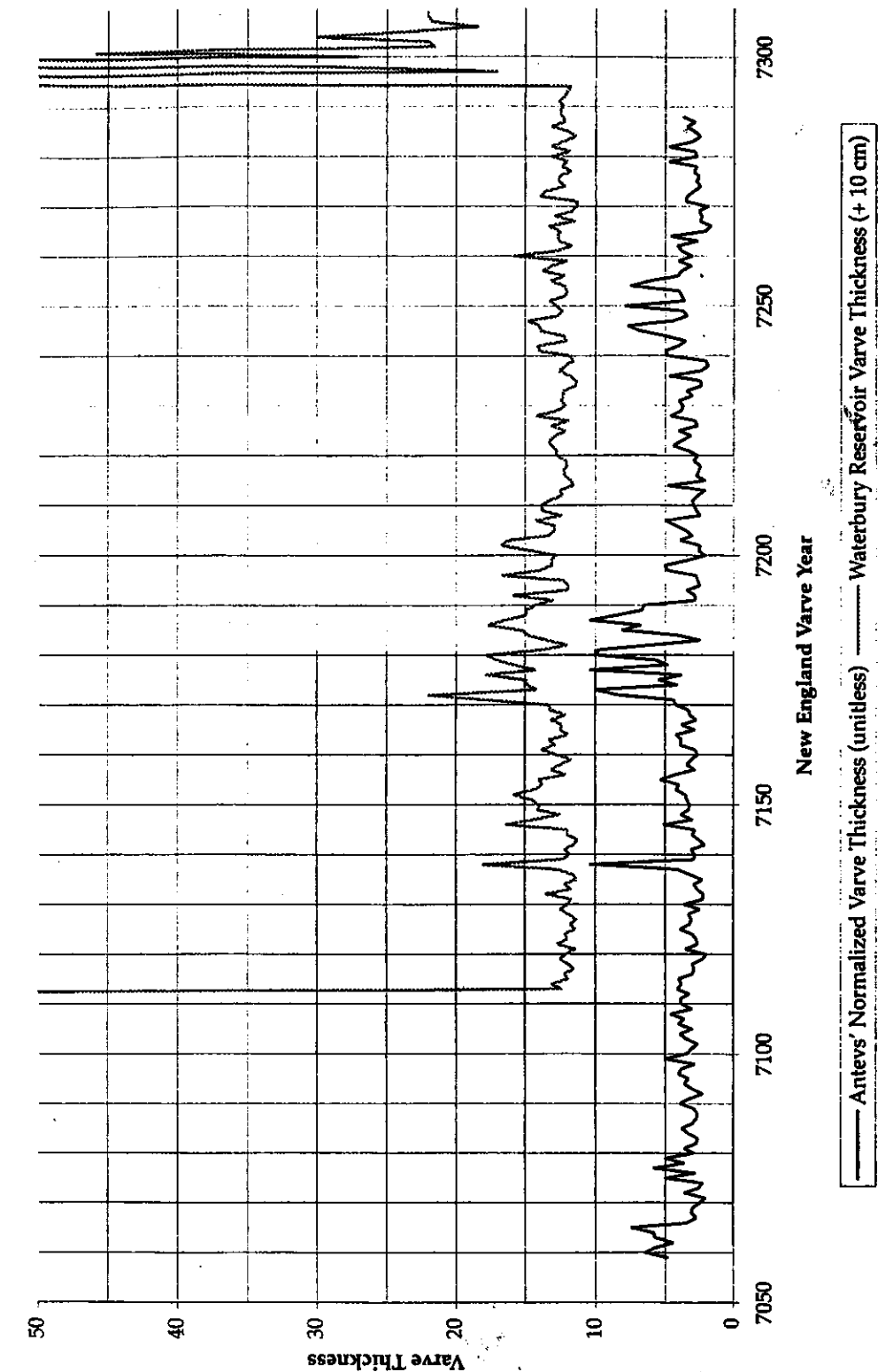


Figure 27: Correlation of varved silt/clay sections from the Waterbury Reservoir with Antevs' (1928) compiled section for the upper Winooski River valley. The thicknesses used in Antevs' curve have been normalized and the units used are unitless. To offset the two curves, 10 cm has been added to the measured thickness of varves in the Waterbury Reservoir section. Thick unit at base of section is a slumped layer. Thick units at the top of the section reflect the draining of Glacial Lake Winooski and the onset of Glacial Lake Mansfield.

Winooski river basin is called Glacial Lake Mansfield (Larsen, 1987b). Locally the elevation of that lake was 690 ft and remnants of the delta produced by Stevenson Brook occur at the Little River campground entrance gate, 300–400 m west of the measured section (Fig. 21). The stratigraphic section indicates that the drop in lake level was rapid and occurred during New England Varve year 7293 or ~13,750 yr BP using the varve correlation chart of Ridge et al. (1999). Beginning at Varve year 7293, silt/clay couplets end and the section is dominated by large influxes of fine, medium, and coarse sand separated by (winter?) silt layers. Bedding between adjacent silt layers is 10's of cm thick and the section contains abundant beds with climbing ripples (Type A and B cross-lamination), indicating current flows to the south or southeast. These sediments are bottom-set beds derived from the newly formed Stevenson Brook delta.

The uppermost stratigraphic unit exposed in the section is a pebble/cobble gravel deposited unconformably on the underlying deltaic facies. The gravel comprises a terrace at eastern end of the peninsula and is most likely old alluvium deposited by the Little River sometime after Glacial Lake Mansfield drained. Alluvial terraces are common adjacent to the reservoir.

Bedding within the stratigraphic section slopes gently east, towards the river valley. Bedding within the upper parts of other stratigraphic sections exposed around the Reservoir all appear to be subhorizontal suggesting that much of the valley now occupied by the Reservoir was filled with glacial lake sediments, now largely removed by the Little River and its tributaries.

Inspection of the currently exposed parts of the Reservoir bottom reveal that almost no historic reservoir sediment has accumulated on the steep slopes. Tree stumps provide further evidence that the old pre-reservoir ground surface is preserved and not accumulating sediment. The deeper and more gently sloping parts of the reservoir bottom were exposed last summer (2002) when the reservoir was drawn down an additional 30 ft to the 520 ft asl level. Historic sediment accumulation over the past 65 years varied from 0 to >3 m and is everywhere rich in organic matter (leaves, woody debris, cones) and historic artifacts (milled wood, trash, and the occasional tent). The contrast between the glacial and historic sediments emphasizes the paucity of vegetation in the recently deglaciated landscape that bordered Glacial Lakes Winooski and Mansfield and the rapid sedimentation rates experienced in that same landscape.

- Leave park, proceed south
- 22.6 Turn left (east) on U.S. Rt. 2
 - 24.0 Turn left (north) on VT Rt. 100
 - 24.2 Turn right, enter I-89 eastbound
 - 27.3 At "Enter Middlesex" sign, look left about 10 ft above the north bound lane to an in-and-out channel/pothole. This was the site of the removal of another pothole during "highway improvement" several years ago.
 - 38.0 Exit 7
 - 45.6 Turn right from I-89 at Exit 5, proceed west on VT Route 64 for 1.6 miles, turn left with caution on steep dirt driveway (now named "Glacier Valley Road"). Proceed about 300 feet past first house (Pollard residence) and park on the right. The descent to Robinson Brook is hazardous and steep and should be undertaken with caution. Property of Kara Pollard (north), Mr and Mrs Paul LaFerriere (middle) and James Cultrera (south).

STOP 11: LaFerriere-Pollard Excavations, Deltaic Beds on Robinson Brook (Larsen)

- (A) The excavation at the LaFerriere residence was mapped on October 1, 2001 (Figs. 28 and 29). Grayish-brown till overlying sheared lake-bottom sediments was exposed on three walls of the excavation. Only till was exposed on the southeast wall. A deeper hole at the north corner of the excavation afforded a 3-dimensional view of the deformed lake-bottom sediments. The drill operator for A & W Well Drilling reported 75 ft of clay and sand on top of bedrock at the LaFerriere well. He used 100 ft of casing and drilled a total of 300 ft for a yield of 8 gallons per minute.
- (B) Excavation for the Pollard residence was mapped at the same time and exposed 6 ft of brown till on 3 walls. The well driller for the Pollard well told the senior author that he encountered clay from a depth of 118 ft to 135 ft with bedrock at 138 ft. However, on the well log he recorded only till from 0 to 138 ft. The well bottomed at 340 ft and had a yield of 4 gallons per minute (Fig. 30).
- (C) Deltaic beds on Robinson Brook were discovered by Norwich students Dan Anderson and Michael Kryposki in the fall of 2001. Deltaic bottomset and foreset beds are exposed in an active slump scar about 60 ft wide and 30 ft high. The bottomset beds consist of flat-lying fine to coarse and pebbly sand. Foreset beds consist of interbedded sand and pebble gravel with dips up to 31° toward S18°W.

- Return to VT Rt. 64, turn right and return to I-89
- 49.0 Turn right (south) on Exit 5 entrance ramp
 - 50.8 Height of land, excellent exposure of Waits River turbidites
 - 61.3 Turn right from I-89 at Exit 4, proceed west on VT Rt. 66
 - 62.9 Turn right at the BLINKER, northwest
 - 63.4 Stop sign, turn right on VT Rt. 12 at the Hodgdon farm

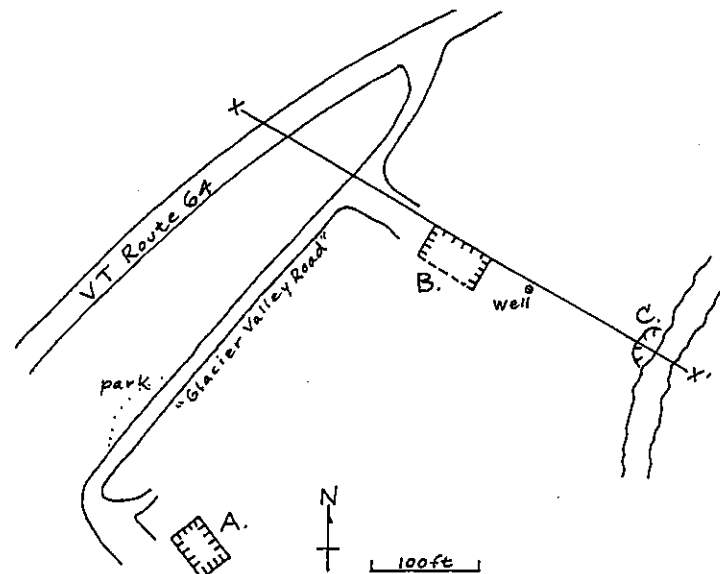


Figure 28. Map of the LaFerriere-Pollard site. A. LaFerriere excavation, B. Pollard excavation, C. exposure of deltaic bottomset and foreset beds on Robinson Brook.

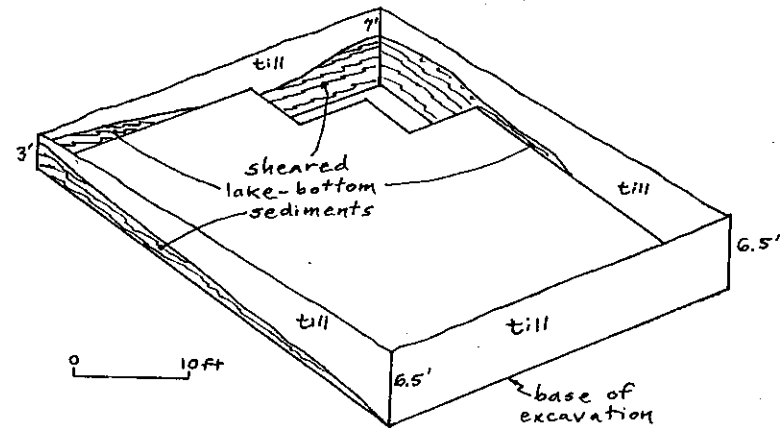


Figure 29. Sheared lake-bottom sediments overlain by till at the excavation for the LaFerriere residence on VT Route 64 1.1 miles from VT Route 12. View looking north.

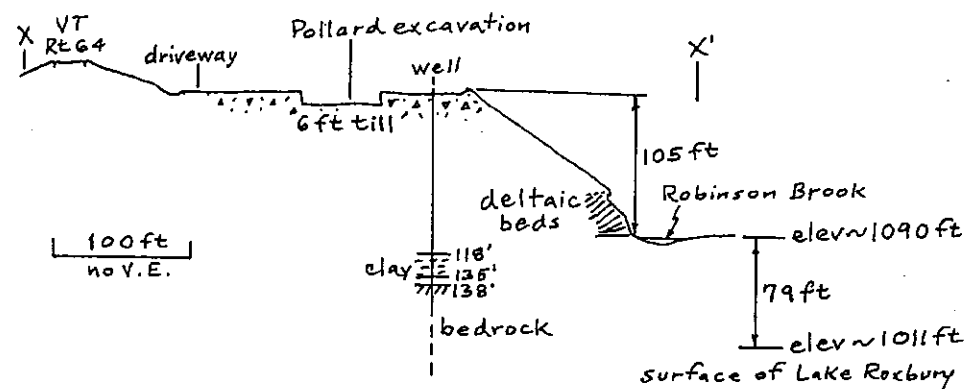


Figure 30. Cross section showing relationship of Pollard site and well to deltaic beds on Robinson Brook.

63.9 Turn left WITH CAUTION ACROSS TRAFFIC to rough farm road, PLEASE do not flatten any more hay than necessary. Stop 13, property of Mr and Mrs Perry Hodgdon

STOP 13: THE ORGANIC SEDIMENTS OF "AYERS POND" (Larsen and Dunn)

During the course of mapping surficial deposits in Ayers Brook valley of the Randolph quadrangle, it became apparent that stinky organic silt was common below stream-terrace deposits (Fig. 31). The silt was very much unlike material in the thick varves of Lake Hitchcock that occur further south in the quadrangle. On November 21, 2001, we discovered and sampled wood and peat-like layers interbedded with fine to very fine sand. The apparent freshness of the wood fragments and the presence of 10-inch burrows in the overlying stream-terrace deposits possibly formed by bank beavers led the senior author to think that we were dealing with a filled-in beaver pond in spite of the fact we found no beaver-cut wood. However, carbon-14 dates of 8700 BP, 8910 BP and 9980 BP took the age of Ayers Pond sediments back to the early Holocene. This suggested that we were dealing with an initial depression that was never filled by Lake Hitchcock sediments, but was blocked by a large subaqueous fan composed of thick varves of Lake Hitchcock at the site of Randolph village (Fig.32). The stream-terrace deposits overlying the organic sediments have a clearly defined fining-upward sequence formed by a meandering stream. Fining-upward sequences are common in terraces at many different levels in the Randolph quadrangle and are an important component of Postglacial history.

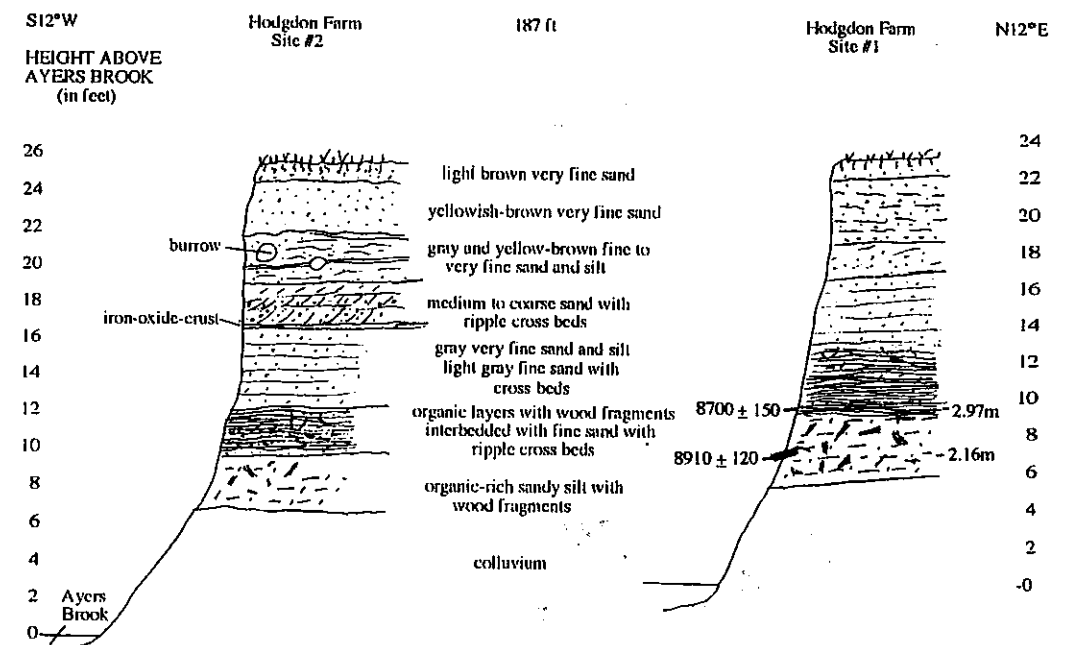


Figure 31. Stop 13. Stratigraphic sections exposed in two cutbanks on Ayers Brook, Hodgdon Farm, Randolph, Vermont. Three separate units in the ponded sediments lie below the iron-oxide crust. Above the crust stream-terrace deposits display a classic fining-upward sequence formed by a meandering stream

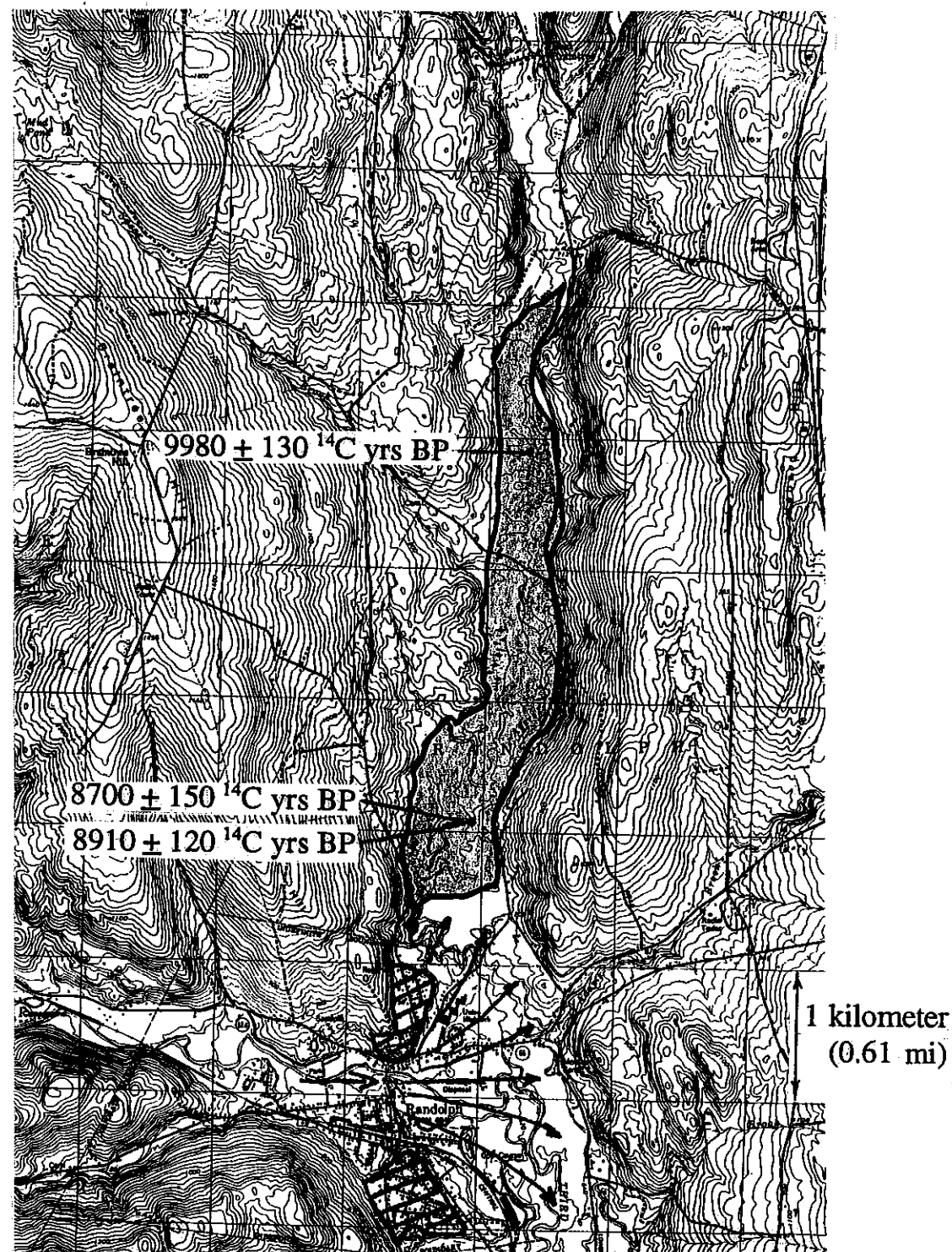


Figure 32. Maximum probable extent of Ayers Pond (heavy line) after the draining of Lake Hitchcock. The pond was dammed by a subaqueous fan (arrows) composed of thick varves that formed as turbidites fed by bottom currents flowing from the west into the north-south valley of Ayers Brook and lower Third Branch. Two areas of terrace/fan deposits that cap subaqueous fan deposits are shown by diagonal lines.

- Proceed south
- 64.5 Hodgdon farm
- 64.7 USGS Ayers Brook gaging station recorded a peak flow of 4,200 cubic feet per second during the flood of June 27, 1998. Based on a 58 yr record, the flow was greater than that in a 100-year flood
- 65.25 Four-way stop, continue straight on VT Rt. 12
- 65.4 Bear left (south) at junction with VT Rt. 12A
- 65.45 Bridge over the Third Branch of the White River
- 65.55 Turn right (west) on School St at the white church
- 66.05 Turn right off School St before crossing railroad tracks
- 66.2 Turn right (north) on Lincoln Avenue, proceed straight through park-like area and descend steep dirt road
- 66.4 Park on gravel areas near house only, PLEASE do not park on grass, owner is Brenda Langevin, walk back up dirt road, turn left at flag for easy access to Stop 14

STOP 14: LINCOLN AVENUE CUTBANK WITH SUBFOSSIL LOGS
(Larsen, Dunn and Springston)

Fourteen subfossil logs are exposed on the south bank of the Third Branch of the White River 0.55 mi, N84°W of the junction of Vermont Routes 12 and 12A in the village of Randolph (Fig. 33). The base of each log is embedded in the middle of point-bar deposits in a terrace that is 12.5 ft above the Third Branch. The 14 logs are exposed over a horizontal distance of 160 ft and are all oriented to the northeast quadrant. The age of one log thought to be a maple is 330 ± 60 ¹⁴C years BP (Geochron No. GX 29694). The stream terrace deposits are underlain by 1.6 ft of fine to very fine sand believed to be Lake Hitchcock thick varves an exposure of which is located at the southeast end of the cutbank. The bottom of the modern channel is an unknown depth below the present surface of the Third Branch making any estimate of the rate of downcutting difficult. However as a minimum, the Third Branch channel has cut down more than 1.6 ft in 330 carbon-14 years.

- End of field trip, thanks for coming! Retrace route, turn left on School St
- 67.3 Turn left (north) from School St, proceed north on VT Rt.12, cross Third Branch
- 67.6 Four-way stop, bear to the right on VT Rt. 66
- 70.1 Exit 4, I-89, take your pick north or south, vans will return to Econo Lodge

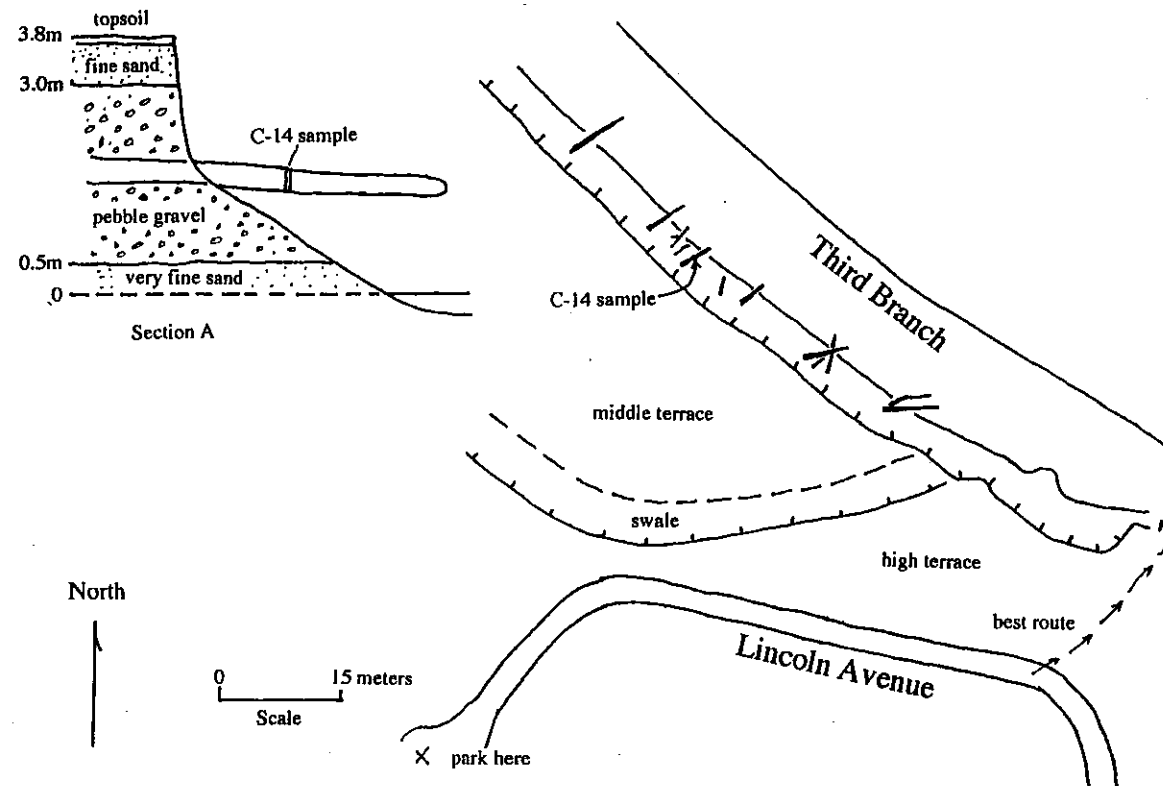


Figure 33. Stop 14. Orientation of subfossil logs in the Lincoln Avenue cutbank 0.6 mi N85°W of the junction of Vermont Routes 12 and 12A in Randolph, Vermont. Maple log in stratigraphic section was dated at $330 \pm {}^{14}\text{C}$ years BP. Thick varves of Lake Hitchcock underlie the high terrace at the far right of the cutbank.

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FRIENDS OF THE PLEISTOCENE REUNIONS
1934-2003

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
1942-45	<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

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

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

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| 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 |
| 66. 2003 | Fred Larsen / Stephen Wright
George Springston / Richard Dunn | Central Vermont |