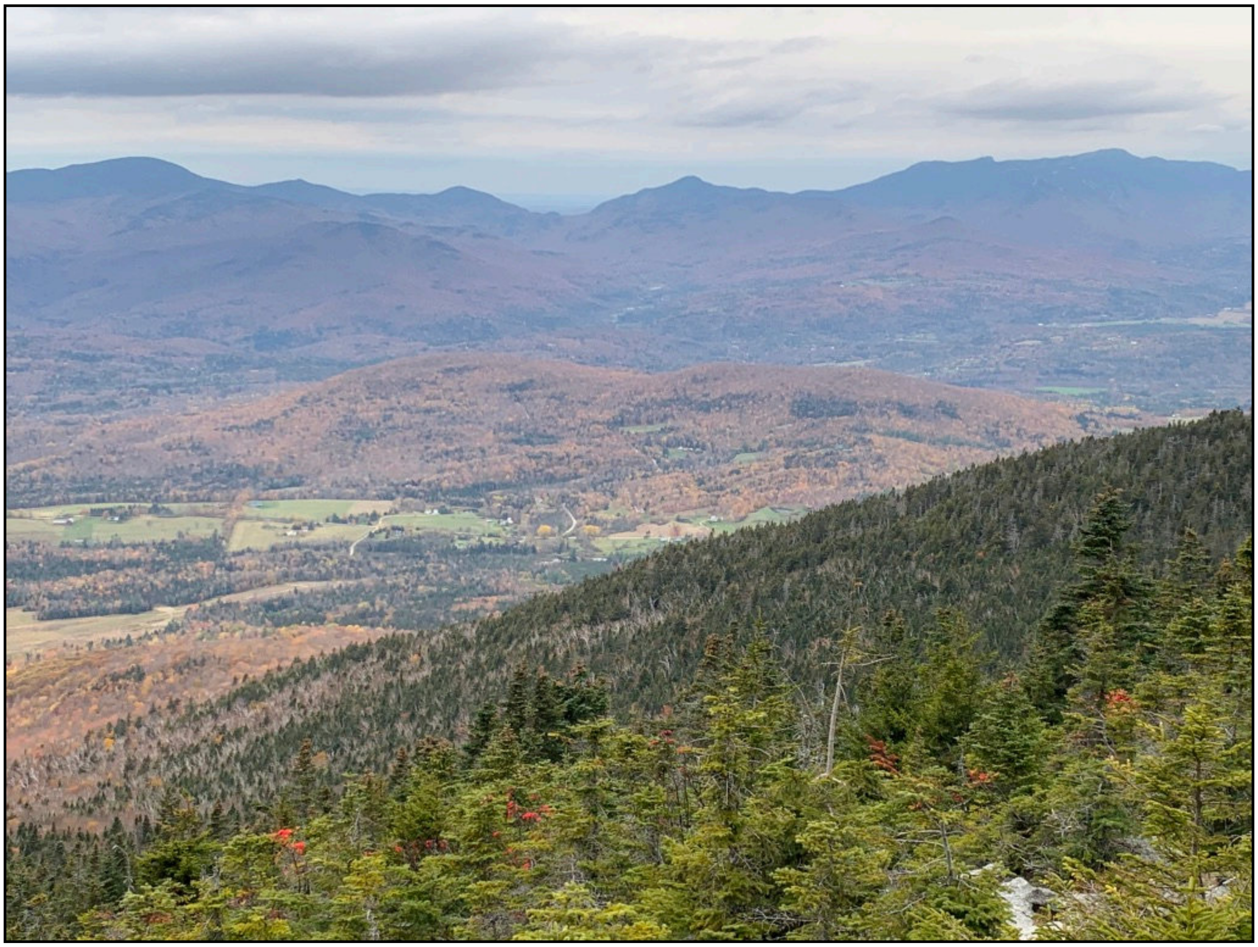
The Glacial, Late-Glacial, and Postglacial History of North-Central Vermont: 20 years of new work

Field Trip Guidebook for the 85th Reunion of the Northeast Friends of the Pleistocene

June 7-9, 2024



View from the top of Mount Worcester looks west towards the Green Mountains. Mount Mansfield, Vermont's highest mountain, is in the upper right of the photo. Nebraska Notch is the low point on Green Mountain ridge line and funneled subglacial meltwater from the Champlain valley beyond the mountains into the valley in the foreground. The whole of the foreground valley was occupied by Glacial Lake Winooski as the Laurentide Ice Sheet retreated northwards (from left to right).

Stephen F. Wright*, Richard K. Dunn, George E. Springston**, Laurie D. Grigg****

*Department of Geography and Geoscience, University of Vermont, Burlington, Vermont
**Department of Earth and Environmental Sciences, Norwich University, Northfield, Vermont

Meeting Itinerary

We welcome you to the 85th annual reunion of the Friends of the Pleistocene, the first Friends' field trip to northern Vermont since 2003. During most of the intervening 20 years our efforts have been directed towards detailed mapping of all 7.5-minute quadrangles within the Montpelier (Vermont) 30' x 60' sheet. Our reunion will celebrate the compilation of the surficial geology across these 32 quadrangles, will highlight our findings, and invite your critique of our work.

Our planned field stops are selected to illustrate (1) glacial/hydrologic processes operating as the ice sheet thinned and retreated across this mountainous terrane, (2) glaciolacustrine processes occurring in the glacial lakes that formed during ice sheet retreat in the regions' valleys, and (3) post-glacial processes and climate changes occurring across the area following the drainage of these lakes. We welcome your critique of our work and look forward to your good company and expertise over the weekend.

Friday Afternoon June 7, 2024: Optional Field trip along the Second Branch of the White River: East Randolph to Williamstown

12:30 PM: Rendezvous at the Middle Branch Market and Deli, East Randolph, Vermont: Field trip by private car.

5:00 PM: Field Trip ends. It's a short drive north to the Berlin Exit on the Interstate where the Comfort Inn is located.

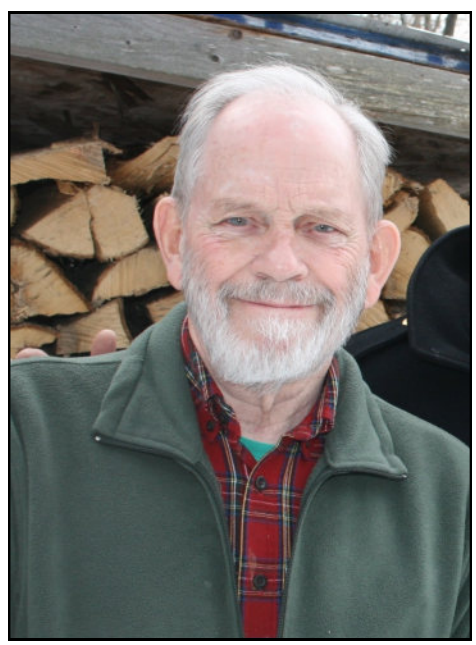
Saturday June 8, 2024: Field Trip in the Stowe/Middlesex Area

7:45 AM: Bus leaves the Comfort Inn, Berlin Vermont

4:45 PM: Bus Returns to Comfort Inn

5:30-7:00 PM: Happy Hour, Banquet at the Barre Granite Museum

7:00 to 8:00: Talk by Laurie Grigg: *From Till to Forest: The evolution of late-glacial climate and vegetation in central Vermont*



Sunday June 9, 2024: Field Trip in the Marshfield/East Barre Area

8:00 AM: Bus leaves the Comfort Inn, Berlin Vermont

2:00 PM: Bus returns to Comfort Inn

Fred Larsen: 1930-2019

All of us were mentored by Fred Larsen, Dana Professor of Geology at Norwich University, and we would like to dedicate this guidebook to him. For many years Fred was one of the few, if not the only, geologist working on the glacial geology of northern Vermont. Fred was a dedicated field geologist and applied his keen mind to many glacial problems, both very local and very regional. Fred was the lead author for the last Vermont Friends' Meeting Guidebook (Larsen et al., 2003a) and we have tried to present work here that he would be proud of. We are grateful for having the opportunity to learn from such a careful, caring, and good natured person.

Table of Contents

Introduction

This field trip guide summarizes our current understanding of the glacial geology of north-central Vermont (Fig. 1). The surficial geology of the entire state was mapped during the 1950's and 1960's and that work compiled to make the "Surficial Geologic Map of Vermont" (Stewart and MacClintock, 1970). During that mapping effort geologists were tasked with mapping 15-minute quadrangles in a single field season which necessitated greater and lesser degrees of reconnaissance mapping. Their mapping efforts were also limited by the detail offered on the 15-minute topographic base maps available at the time. On the other hand, they had the advantage of working in a landscape that was more open and accessible and at a time when many active pits were supplying sand and gravel for interstate construction, many of which have been long closed. Also assembled at this time was an inventory of highway construction materials where the state surveyed every town to map the extent of sand and gravel deposits available for future highway construction projects (see Vermont Geological Survey website for downloads of these reports).

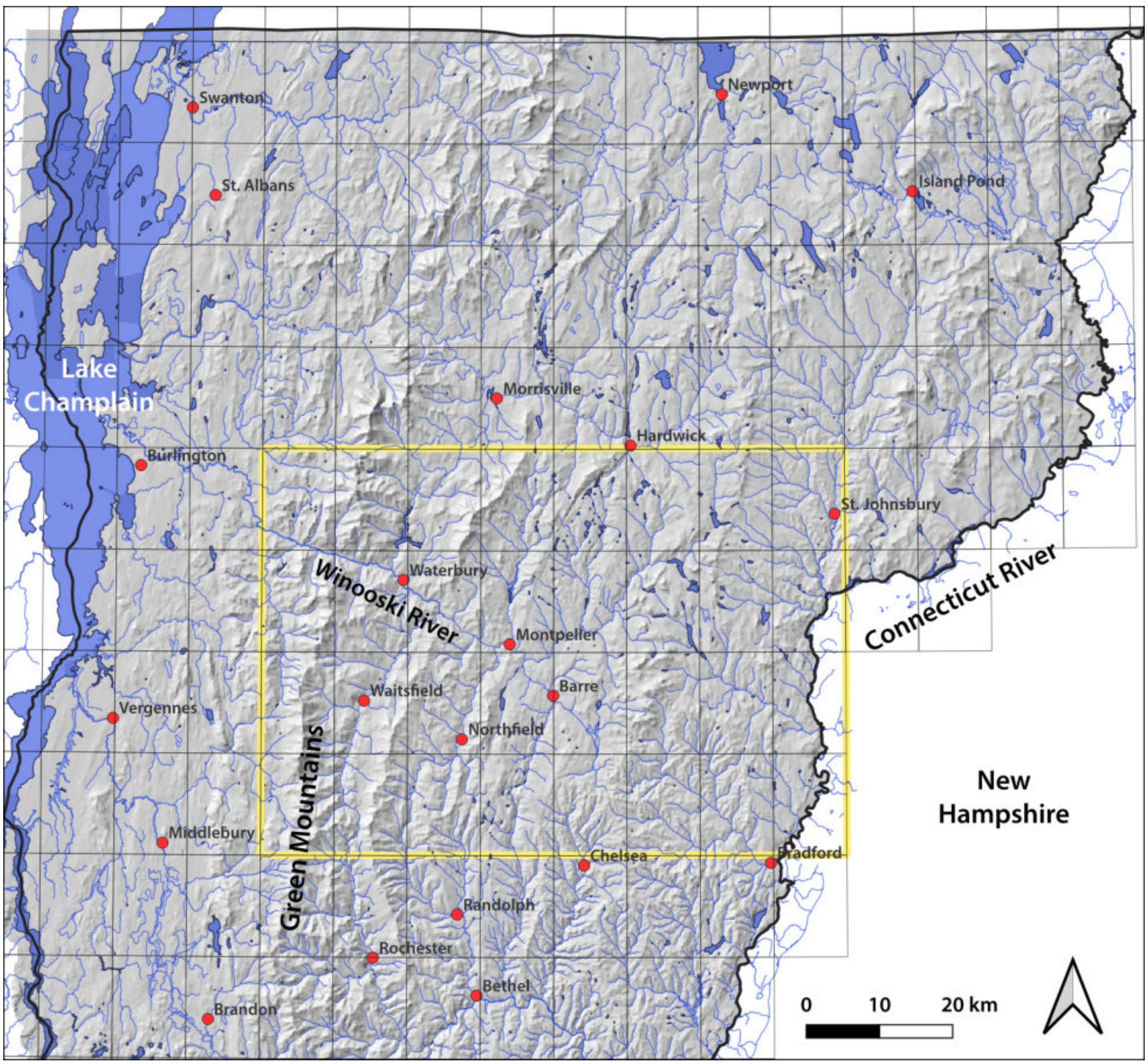


Figure 1: Physiographic map of northern Vermont. Yellow box outlines the extent of the Montpelier 1:100,000 sheet, the focus of this field trip. Smaller boxes are individual 7.5-minute quadrangles.

The impetus for this gathering of the Friends of the Pleistocene is the completion of detailed mapping of all 32 7.5-minute quadrangles within the Montpelier 1:100,000 sheet (Fig. 1) . This mapping work began in the 1990's using paper quadrangle maps and pencils and evolved to utilize the latest phone-based, GPS-enabled mapping software on detailed LiDAR shaded-relief base maps. This mapping work has led to a better understanding of a wide variety of different glacial processes occurring as the Laurentide Ice Sheet flowed across the region at the Last Glacial Maximum and then subsequently thinned and retreated from the area. We outline below many of the ideas presented over the last 20 years in GSA talks, Vermont Geological Survey Reports that accompany individual quadrangle maps, and NEIGC field guides. The following sections of this field guide provide background to the field stops which are described as the last part of this guide.

Ice Flow History

Glacial till, erratics, and striations on the summits of the region's highest mountains indicate that the ice sheet in northern New England was sufficiently thick to completely cover the mountains. The direction of ice flow is interpreted from glacial striations, rouches moutonnées, erratic dispersal fans, and most recently by crag and tail and streamlined till landforms visible in LiDAR imagery. While till fabrics were utilized extensively in the 1960's to interpret several different regional ice flow directions occurring in separate tills (Stewart and MacClintock, 1969, 1970), this work has been largely discredited (see discussion in Wright, 2015b).

Using erratic dispersal fans emanating southeast and south from intrusive granite stocks in central Vermont, Larsen (1972) concluded that the Laurentide Ice Sheet had only flowed towards the southeast and south during the late Wisconsinan. Larsen (2003b) was also able to successfully trace a train of distinctive gneissic erratics from central Vermont in a NNW direction to their source, the St. Didace pluton in the Laurentian Mountains of Québec. A recent map showing the locations of distinctive orbicular granite erratics emanating from the one known outcrop in Craftsbury shows a very similar dispersal fan (Fig. 2).

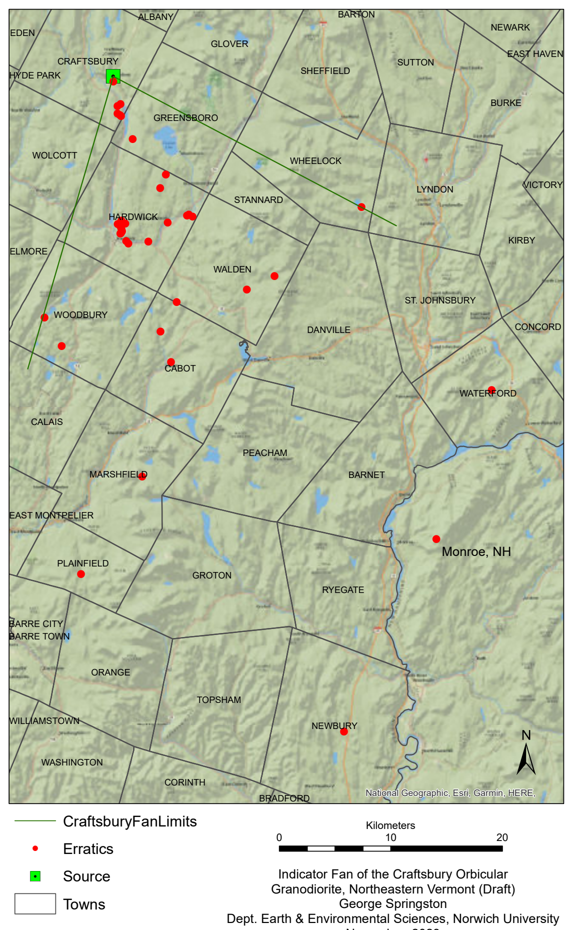


Figure 2: Erratic dispersal fan of Craftsbury orbicular granite.

A compilation of glacial striations observed in northern and central Vermont was published by Wright (2015b) and this compilation has been updated with new data to accompany publication of new Montpelier Surficial Geologic Map (Fig. 3). The most widely distributed set of glacial striations are aligned generally northwest-southeast across northern Vermont and are particularly well preserved along the high ridge lines of both the Green Mountains and lower elevation mountain ranges to the east (Northfield, Worcester, and Lowell Ranges, Fig. 3, 4). This striation set maintains a relatively uniform azimuth across north-central Vermont and is parallel with striations measured across a wide swath of south-central New Hampshire and southern Maine, areas immediately down-glacier from north-central Vermont (Goldthwait in Antevs, 1922; Thompson and Borns, 1985) as well as in adjacent parts of southern Québec (Parent and Occhietti, 1999). The regional distribution of this striation set, particularly at high elevations, suggests a broad area of southeast ice flow existed at the LGM with ice flowing obliquely across the mountains in northern Vermont.

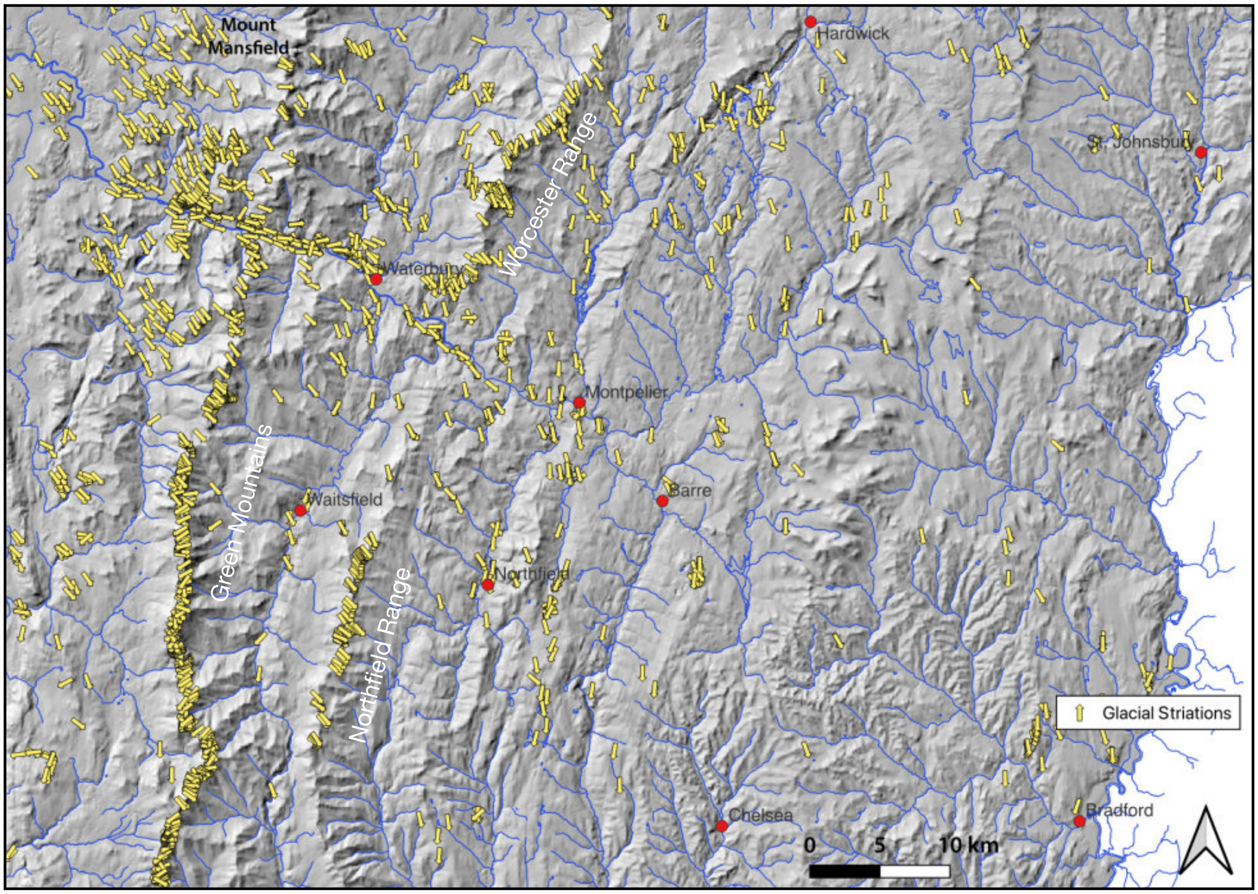


Figure 3: Compilation of glacial striations across north-central Vermont. Regional ice flow to the southeast is recorded by striations along mountain ridge lines. Younger, topographically controlled ice flow, is evidenced by striations parallel to valleys. A limited area of southwest ice flow across the southern Green Mountains records thinning of Champlain valley ice, possibly a result of ice streaming in the valley.

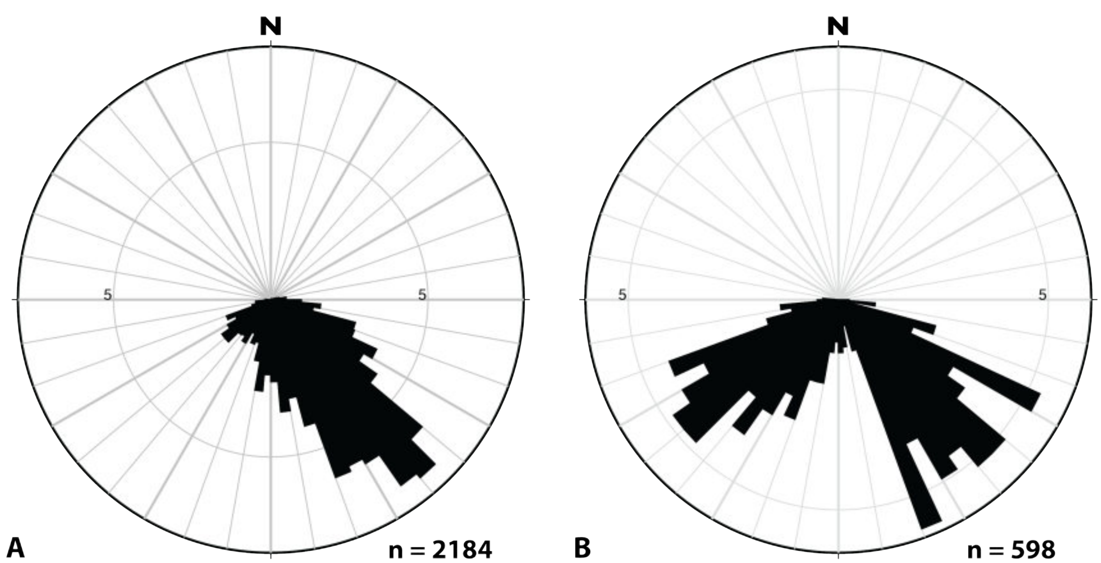


Figure 4: Rose diagram plots of down-glacier striation azimuths. Data are subdivided into 5° increments. Radial scale is percentage of all striations plotted. (A) Plot of all 2184 striations recorded in the field area. The strong southeast maxima overwhelms lesser numbers of striations with southerly and southwesterly azimuths. (B) Plot of all 598 striation azimuths measured along and adjacent to the ridgeline of the Green Mountains south of the Winooski River valley. The strongly bimodal distribution is clearly evident with 78% of all plotted striations aligned either southwest (39% between 200 and 250) or southeast (39% between 115 and 160).

Figure 5 models the ice sheet profile across New England at a time when the ice sheet was flowing to the southeast at or near its maximum extent. The ice sheet profile is computed using the ground surface slope and a spreadsheet algorithm developed by Benn and Hulton (2010) set to maintain a uniform basal shear stress of 100 kPa, an average value. The calculated ice sheet surface is well over 2 km above the Green Mountain ridgeline and almost 3 km above the ground surface in the Champlain valley (right side of Fig. 5). Lower basal shear stresses would thin the modeled ice sheet profile whereas higher basal shear stresses would thicken the modeled profile. It should be noted that in contrast to lower elevation areas, exposure age dates from the highest elevation peaks in New England have consistently recorded old dates indicating relatively little bedrock was eroded from these peaks by the ice sheet, possibly because the ice sheet was cold-based in these areas (Bierman et al., 2015; Corbett et al., 2019; Halsted et al., 2022). The below model does not account for these "sticky spots" at the base of the ice sheet and, given their very small area across New England, they are unlikely to significantly affect the ice sheet profile.

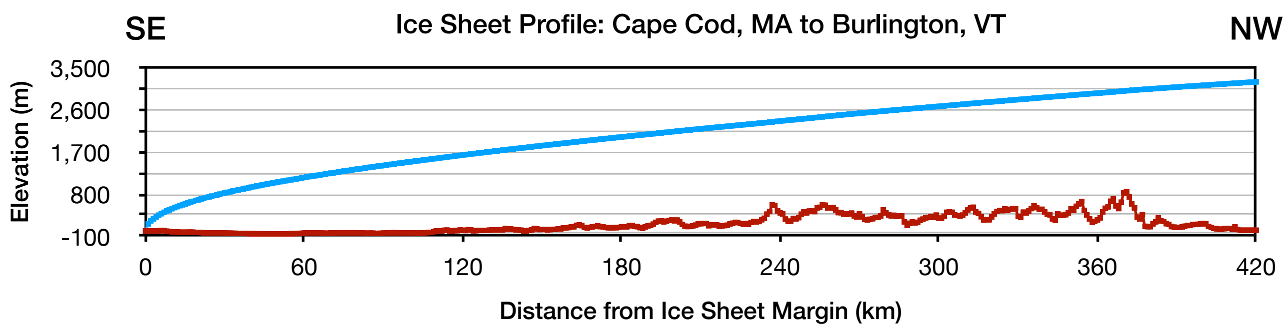


Figure 5: Profile of modeled ice sheet surface and New England topography between Cape Cod, Massachusetts and Burlington, Vermont. Ice sheet model maintains a constant basal shear stress of 100 kPa. Between 2 and 3 km of deforming glacial ice lay above the Green Mountains and adjacent valleys. Vertical Exaggeration ~18X.

In the many mountain valleys the youngest striations usually parallel those valleys (Fig. 3). The logical interpretation is that topography began influencing ice flow as the ice sheet thinned, a common observation along the margin of modern ice sheets. In most areas ice flow shifted from southeast to south as most of the valleys in the region are aligned parallel to the north-south trend of the mountains (Fig. 3). This is particularly evident in the Champlain valley west of the mountains and in the broad area east of the mountains centered on the Connecticut River valley. Large scale exceptions to this are the Lamoille and Winooski river valleys which both cut completely across the mountains and effectively funneled ice from the Champlain Valley ESE across the mountains. On the west side of the mountains glacial striations converge into these valleys and the aerial extent of this convergence is an indication of the area of ice in the Champlain Valley that was contributing to these outlet glaciers (Fig. 3). A fuller description of the glacial lake history follows in a later section, but these outlet glaciers were responsible for damming those major valleys cutting across the mountains producing large glacial lakes east of the mountains (Larsen, 1972, 1987).

From Appalachian Gap south to Killington, a younger set of striations occurs along the Green Mountain ridge line and is oriented from northeast to southwest (Figs. 3, 4). These striations indicate that an abrupt shift in ice flow occurred from southeast to southwest resulting in ice now flowing across the mountains into the Champlain valley (Ackerly and Larsen, 1987; Wright, 2015b). Wright (2015b, 2017) attributed this change in ice flow to the initiation of an ice stream in the southern Champlain and northern Hudson River valleys. Abundant north-south striations, crag and tail structures, and streamlined bedforms with length to width ratios exceeding 10:1 within the Champlain valley all indicate that fast north to south ice flow occurred parallel to this large-scale topographic trough (Wright, 2015b).

The Evolution of Glacial Lakes in the Winooski River Drainage Basin

The physiography of much of central-northern Vermont is dominated by the watersheds of the westward flowing Winooski and Lamoille rivers as they cut perpendicularly across the NNE-SSW strike of folded, early to middle Paleozoic metasedimentary rocks (Fig. 1). Major tributaries to the Winooski and Lamoille rivers flow parallel to bedrock strike. In this section we outline the evolution of glacial lakes in these valleys concentrating on areas east of the Green Mountains. We note, however, that our work in this region has not fundamentally altered the glacial lake history first sketched by Merwin (1908) and later deciphered in some detail by Larsen (1972, 1987). Larsen (1972, 1987) also presents excellent reviews of early work in the Winooski drainage basin leading to our current understanding of this history.

During glacial retreat across central Vermont the ice margin dammed the north-draining tributary valleys of the Winooski; from west to east the Mad River, the Dog River, and the Steven's Branch. These valleys filled with meltwater and the subsequent lakes drained over southern thresholds (Fig. 6; Larsen, 1972, 1987). Deltas located in each of these valleys are graded to the different lake levels, confirming the relative elevations of these lakes (Dunn et al., 2007; Larsen, 1972, 1987).[1]

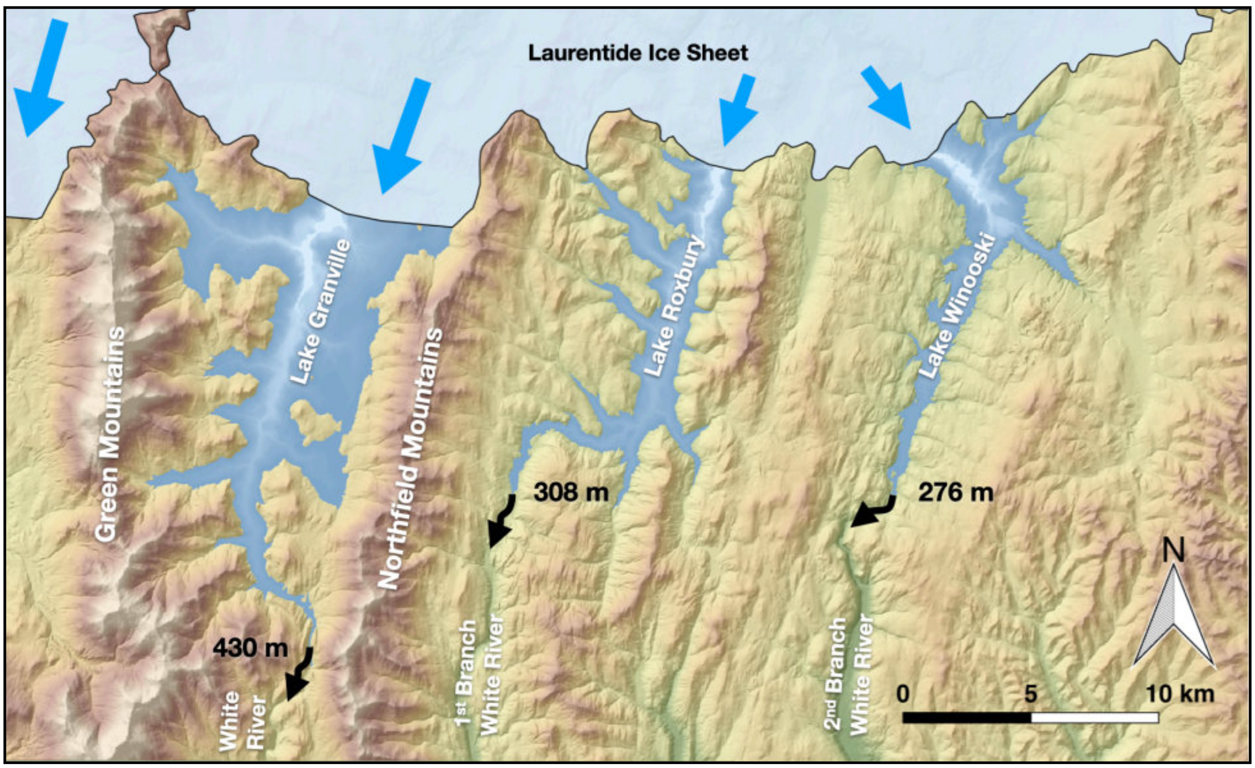


Figure 6: The retreating Laurentide ice sheet east of the Green Mountains dammed lakes in north-flowing tributary valleys. Glacial Lake Granville occupied the Mad River valley, Glacial Lake Roxbury the Dog River valley, and Glacial Lake Winooski (originally named Glacial Lake Williamstown by Larsen (1972, 1987)), the Stevens Branch valley. Blue arrows show directions of ice sheet flow.

These early lakes grew in size as the ice continued retreating northwards until the ice margin approached the low-lying Winooski River valley allowing the small lakes to coalesce and form a single larger lake, Glacial Lake Winooski, with a spillway across the lowest 276 m drainage divide (Fig. 7; Larsen, 1972, 1987). Recent mapping of this spillway area has shown that the lake outlet elevation originally identified by Larsen (279 m (915 ft); 1972, 1987) is covered by

[1] Glacial lake projections are largely from Springston et al. (2020) and use an isostatic tilt of 1.15 m/km to N21W.

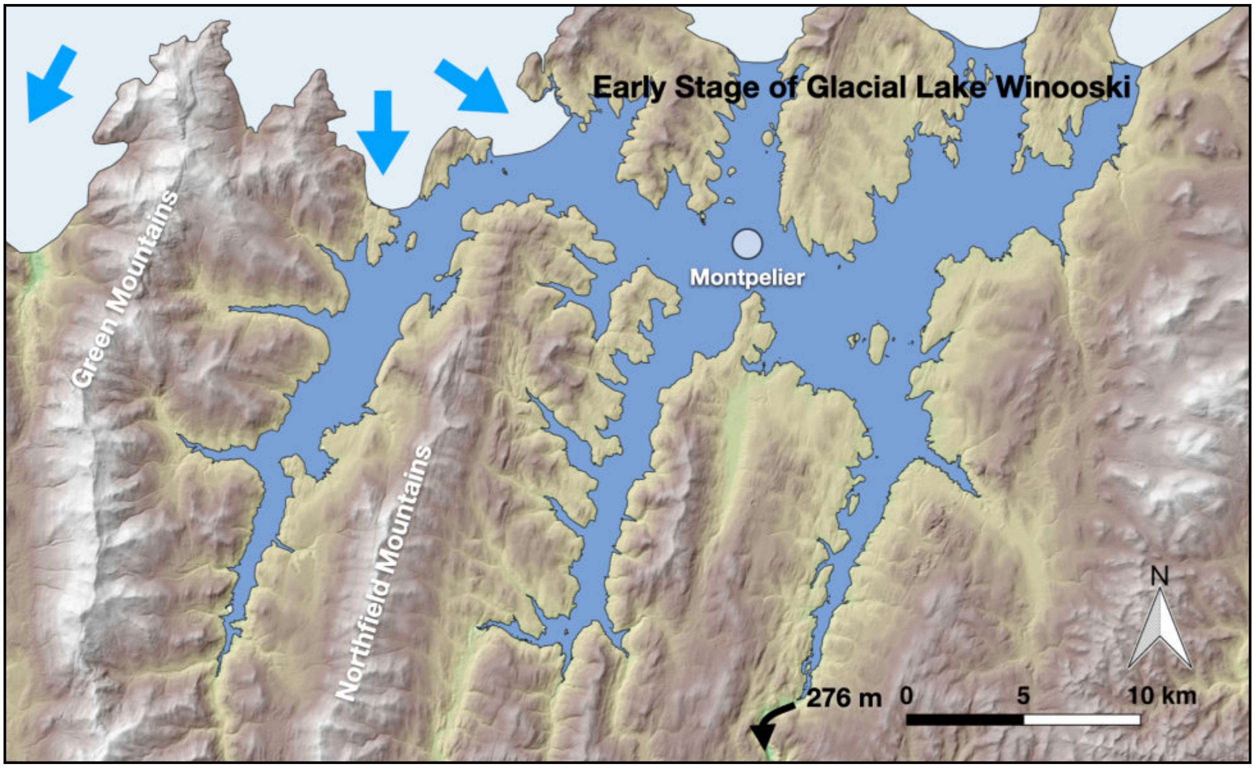


Figure 7: Glacial Lake Winooski greatly expanded when northward retreat of the ice sheet margin allowed the small lakes in the Mad, Dog, and Stevens Branch valleys to merge and utilize a common outlet through Williamstown Gulf at 276 m. Blue arrows show inferred ice flow directions.

two Holocene alluvial fans and the outlet elevation is actually a few meters lower in elevation (276 m) and several hundred meters farther south (Wright, 2022b). The lake continued expanding north following the retreating ice front and eventually inundated both the Winooski and Lamoille River drainage basins (Fig. 8). The maximum northwestern extent of the lake is known from the limit of Glacial Lake Winooski sediments mapped in the Lamoille River valley (Wright, 2001). Drainage of the lake remained at the Williamstown Gulf spillway in the southeast. During this time the water discharging over that spillway was sourced not only from meteoric water across both drainage basins, but glacial meltwater from rapidly thinning and retreating ice east of the Green Mountains and meltwater routed from the Champlain valley across the mountains via the Lamoille and Winooski River valleys. No discharge estimates have been made, but considerable volumes of water must have been flowing across this spillway.

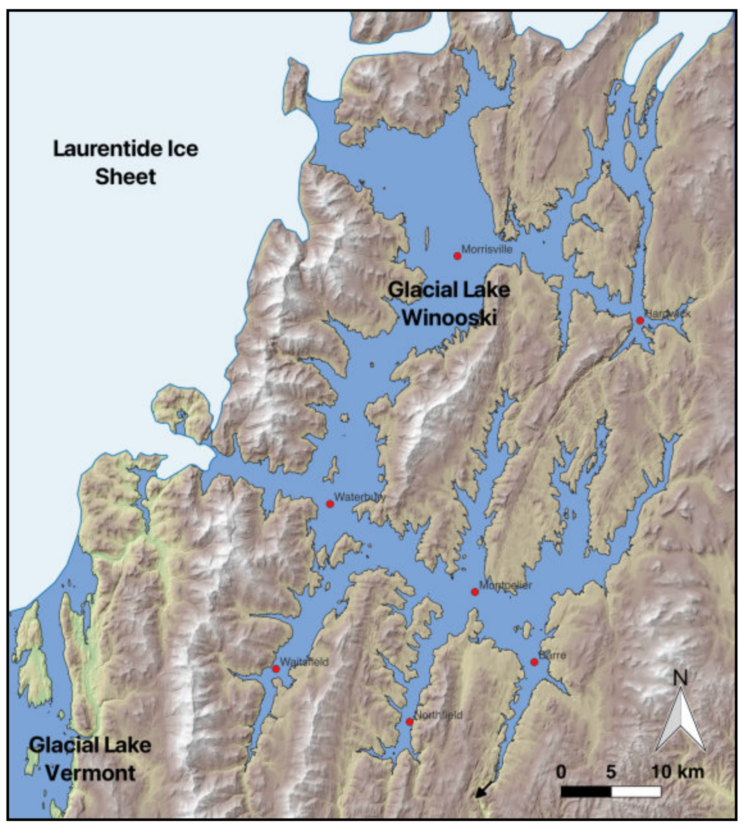


Figure 8: Maximum extent of Glacial Lake Winooski in the Winooski and Lamoille River basins. The lake existed as long as ice in the Champlain valley dammed these WNW-draining valleys.

Continued ice retreat down the Winooski valley eventually opened two new closely spaced outlets in the vicinity of

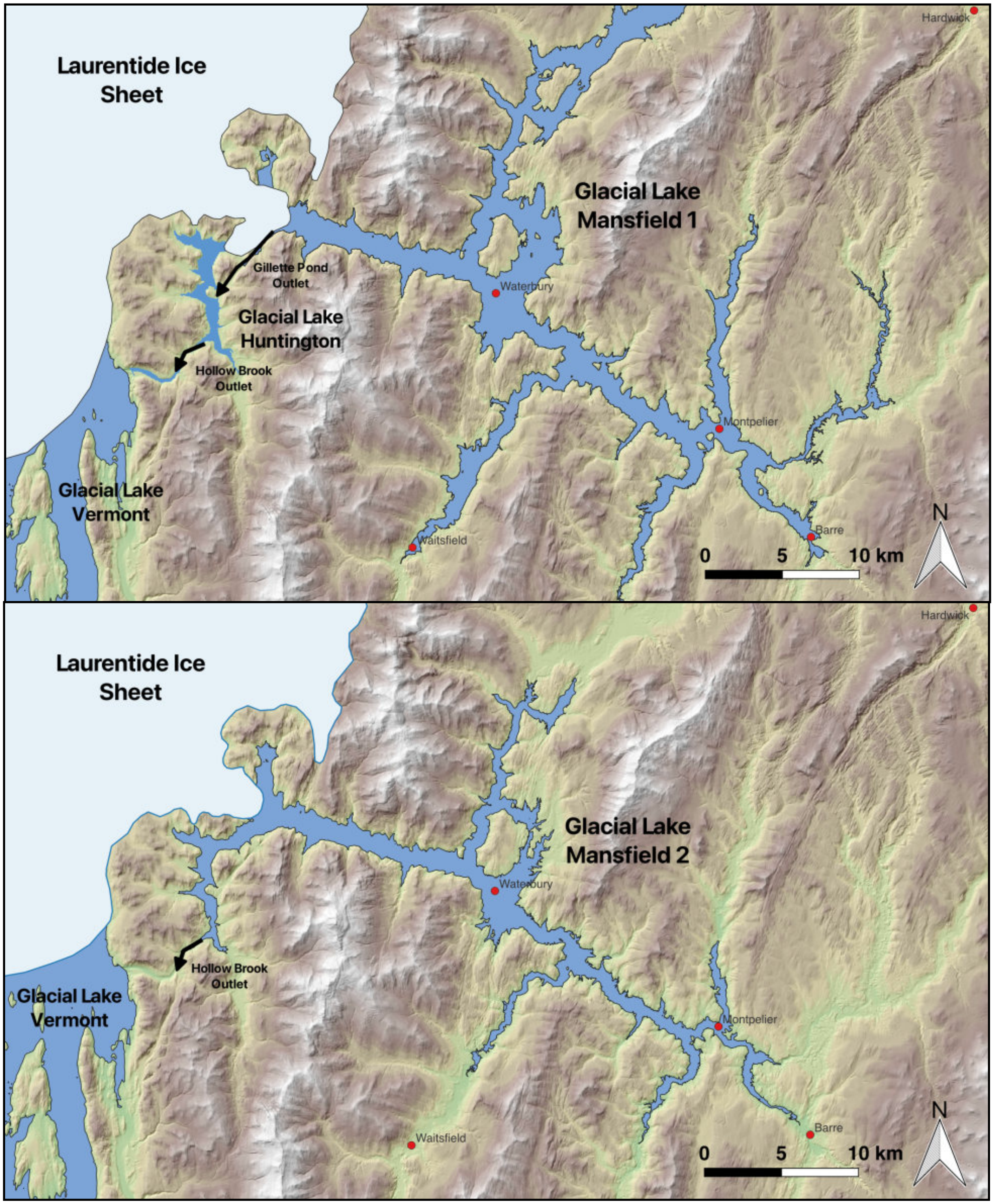


Figure 9: Evolution of Glacial Lake Mansfield in the Winooski River basin. Glacial Lake Mansfield 1 formed when the ice sheet retreated WNW down the Winooski River valley far enough to uncover the Gillette Pond outlet rapidly lowering water elevations ~90 m. Relatively little further retreat allowed the lake to drop another 26 m and merge with Glacial Lake Huntington to form Glacial Lake Mansfield 2. Both lakes drained via the Hollow Brook outlet into Glacial Lake Vermont.

the Huntington River (Fig. 9). The first of these, the Gillette Pond outlet, lowered Glacial Lake Winooski by ~90 m rapidly draining ~375 km^3 of water from the lake basin (Wright, 2018a). It's unclear whether the ice dam failure was via (1) a subaerial breach (water flowing over the ice sheet), (2) a subglacial tunnel, (3) flotation of the ice margin, or some combination of these processes. Regardless, the ice dam failure was likely catastrophic as friction from water leaking out of Glacial Lake Winooski rapidly melted the ice sheet surrounding the outlet passageway. The "new," lower-elevation lake is referred to as Glacial Lake Mansfield 1 which drained, via Glacial Lake Huntington in the Huntington River valley into Glacial Lake Vermont (Fig. 9; Larsen, 1972, 1987). A well-preserved delta and channel occurs where the Gillette Pond outlet enters the Huntington River valley (Fig. 10).

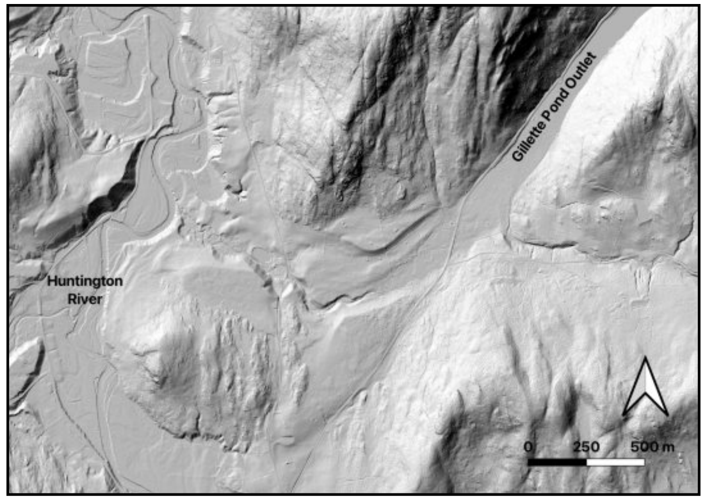


Figure 10: Delta and channel formed where water flowing through the Gillette Pond outlet entered Glacial Lake Huntington. See Figure 9 for location.

Only several hundred meters of additional ice retreat allowed Glacial Lakes Mansfield and Huntington to merge and utilize a common outlet at the Hollow Brook threshold ~26 m lower than the Gillette Pond outlet (Fig. 9). This lower elevation stage is referred to as Glacial Lake Mansfield 2 (Larsen, 1972, 1987).

When the ice sheet retreated into the Champlain valley, Glacial Lake Mansfield 2 drained through an outlet in Williston and the Winooski River valley became an arm of Glacial Lake Vermont (Fig. 11). Glacial Lake Vermont underwent several significant drops in elevation during its history (Chapman, 1937; Rayburn et al., 2011; Rayburn et al., 2005; Van Hoesen et al., 2016), but the shoreline changes associated with these different lake stages in the Winooski River valley are limited due to the steep topographic gradients (Fig. 11).

Typical of areas inundated by glacial lakes, large parts of the Winooski River basin below the elevation of Glacial Lake Winooski are blanketed with a variety of glaciolacustrine sediments (Wright et al., 2023a). In most areas these sediments occur in a fining-up sequence that records the retreat of the ice front, the source of most of the sediments, progressively farther from the locus of sediment deposition (Larsen, 1987). The sections typically begin with coarse esker sediments which transition into subaqueous fan deposits (aka ice-proximal deposits). These then transition to thick sequences of very fine and fine sand and then to varved silt and clay, i.e. ice-distal deposits. Our mapping indicates that in the shallow arms of Glacial Lake Winooski rapid sedimentation adjacent to the retreating ice front likely completely filled those parts of the lake with sediments. The significant ~90 m drop in lake level from Glacial Lake Winooski to Glacial Lake Mansfield 1 triggered rapid erosion of earlier-deposited sediments and their later redeposition as deltas along the margins of the lower-elevation lake.

During deglaciation, many small, unnamed proglacial lakes formed in the irregular topography of central Vermont. Similarly, during a short-lived and probably localized readvance of the ice margin, small proglacial lakes were impounded in several valleys. On Sunday we will visit depositional sequences in two of these lakes, glacial Lake Nasmith and glacial Lake East Barre (Dunn and Springston, 2019), which formed in north-flowing tributaries of the upper Winooski River.

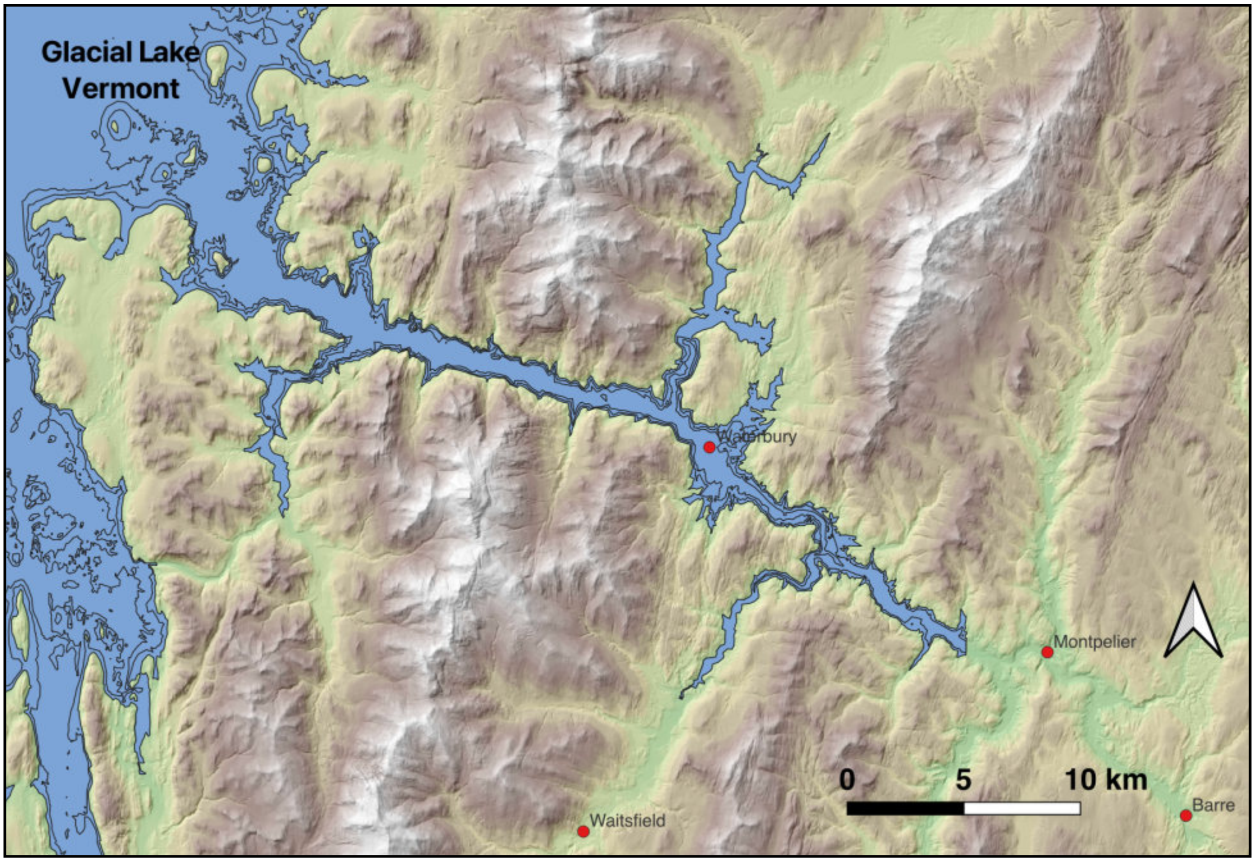


Figure 11: The Coveville stage of Glacial Lake Vermont flooded much of the Winooski River valley. Black lines within the Coveville lake mark both the upper and lower Fort Ann Stages of Glacial Lake Vermont.

Timing of Ice Advance and Retreat Across the Region

The timing of ice advance across New England has been difficult to asses. Based on the youngest OSL ages of cave sediments in the Champlain valley Munroe and others (2016) concluded that the Laurentide Ice Sheet did not advance south of the St Lawrence lowlands into the Champlain valley until early in MIS-2, ~35 ka BP. The ice sheet subsequently advances across New England reaching the terminal moraine by ~25 ka BP (Balco et al., 2002; Corbett et al., 2017) a time span of ~10,000 years.

Our understanding of the timing of ice sheet retreat across northern Vermont comes from both the glacial lake varve record (the North American Varve Chronology; Ridge et al., 2012) and from exposure age dating. We begin with the general observation that the ice sheet east of the Green Mountains retreated northward much earlier than the ice sheet in the Champlain valley. We know this because Glacial Lake Winooski expanded well north of the Lamoille River valley at the same time that ice much farther south in the Champlain valley was sufficiently thick to occupy and dam the Winooski River forming the lake (Fig. 8).

Building on early work by Antevs (1922, 1928), Ridge and others (2012) successfully correlated and dated numerous sections of varves along the Connecticut River valley establishing the North American Varve Chronology (NAVC) and dating the retreat of the ice sheet up the valley. Figure 12 transcribes Ridge's (2016) dated ice margin positions along a portion of the Vermont/New Hampshire border. During this period of time the ice sheet was retreating at a rate of ~300 m/yr, markedly faster than its 80-90 m/yr retreat rate across Massachusetts' border with Vermont and New Hampshire (Ridge et al., 2012).

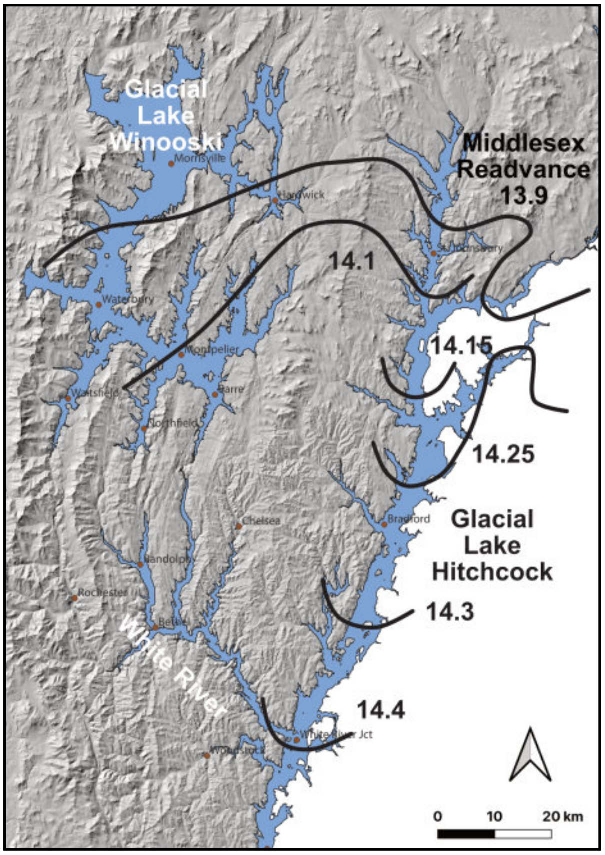


Figure 12: Map shows the extent of Glacial Lakes Winooski and Hitchcock across northern Vermont and New Hampshire. Dark lines showing successive positions of the retreating ice front are taken from Ridge (2016). Numbers are ages in calibrated kyr BP based on the North American Varve Chronology.

The increase in both grain size and layer thickness occurred when the ice dam in the Winooski Valley retreated far enough down-valley (WNW) for water in Glacial Lake Winooski to rapidly escape to the Champlain valley via the Gillette Pond outlet in Huntington, Vermont (Fig 9; Larsen, 1972, 1987; Larsen et al., 2003b; Wright, 2018a). The rapid drop in lake level associated with this breakout flood allowed much coarser sediment (derived from shore) to reach the site of the Waterbury Reservoir section.

The beginning of Glacial Lake Winooski is harder to date because no good sections of varved lake sediments have been found in the Stevens Branch valley, the valley hosting the lake during its earliest years immediately north of the lake's Williamstown Gulf outlet (Figs. 7, 14).

The extension of the 14.1 and 13.9 kyr ice margin positions across the Winooski drainage basin is based on the correlation of Glacial Lake Winooski varve sections originally measured by Antevs (1928; Fig. 13). More recently, three new sections of varved Glacial Lake Winooski sediments have been measured and a fourth, originally mapped by Antevs, has been remeasured and extended (Fig. 13). Specifically these sections are the Muzzy Brook (Larsen et al., 2001), Montpelier (Springston and colleagues, unpublished), Wrightsville Reservoir (Wright and students, unpublished), and Waterbury Reservoir Sections (Wright, in Larsen et al., 2003b). Figure 14 shows how these sections are correlated with Antevs' (1928) Winooski Valley varve compilation and the North American Varve Chronology.

The Waterbury Reservoir section (Fig. 14, red line) clearly records when Glacial Lake Winooski partially drained. An abrupt change from silt/clay varves to thick fine sand/silt layers occurs at North American Varve 6976 which was deposited ~13,844 yr BP based on dating of the North American Varve Chronology[2] (Thompson et al., 2017).

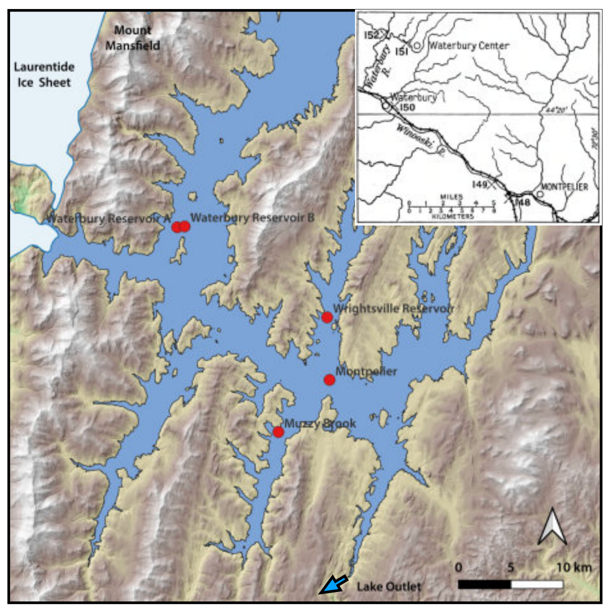


Figure 13: Locations of measured varve sections within the projected extent of Glacial Lake Winooski. Inset is Antevs' (1928) map showing the locations of varve sections he measured (148-152). Of Antevs measured sections, only 152 is still exposed (Waterbury Reservoir A). Mount Mansfield, the highest mountain in Vermont, is the location of exposure age dates by Corbett and others (2019).

[2] Calibration of the NAVC has an uncertainty of at least 100 years (Thompson et al., 2017).

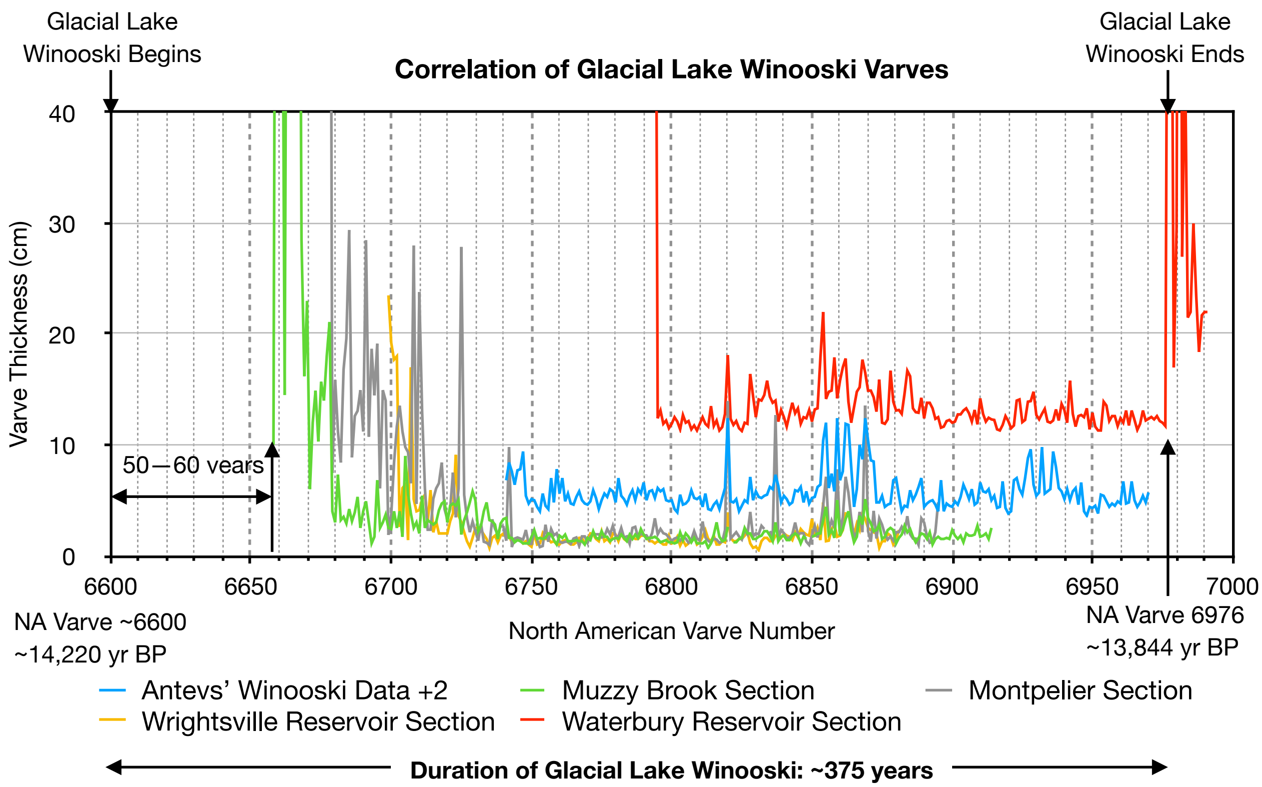


Figure 14: Correlation of 4 measured sections of Glacial Lake Winooski varves with Antevs' Winooski compilation (see Fig. BB for location of sections). The Waterbury Reservoir section records the partial draining of Glacial Lake Winooski at NA Varve 6976. The Muzzy Brook section contains the oldest ice-proximal varves which were deposited 50-60 years after the ice sheet crossed the Glacial Lake Winooski outlet (see text for details).

However, the Muzzy Brook, Montpelier, and Wrightsville Reservoir varve records provide the basis for a good estimate. While only the Wrightsville Reservoir section begins directly above till, the Muzzy Brook and Montpelier sections both begin with thick, ice-proximal varves and likely record deposition shortly after the ice sheet retreated north of these respective locations. From the varve correlation graph (Fig. 14) it's clear that sediments begin accumulating first at the southernmost section (Muzzy Brook, green line), begin next in Montpelier (grey line), and last at the northernmost section (Wrightsville Reservoir, orange line). Varves thin over a span of 20 to 30 years as the ice sheet, the source of most of the sediment, retreats farther away (Fig. 14).

Ice margin retreat rates can be calculated between Muzzy Brook, Montpelier, and the Wrightsville Reservoir using the time interval between the start of sedimentation at each site and the distance between these sites (Fig. 15,A,B). The calculated retreat rates between Muzzy Brook and Montpelier (295 m/year) and Montpelier and the Wrightsville Reservoir (305 m/year) are almost identical. Furthermore, Ridge and others (2012), using many more sections of varved lake sediments, calculate an identical retreat rate of ~300 m/year in the upper Connecticut River valley, the reach of the valley due east of the graphed sections (Fig. 12).

If the ice retreated north from the outlet of Glacial Lake Winooski to the latitude of the Muzzy Brook section at 300 m/year it would have taken ~50 years to retreat that distance (Fig. 15, D). A similar calculation based on a more conservative retreat rate of 250 m/year increases the duration of that retreat to 60 years (Fig. 15, E). These calculations indicate that the ice margin retreated across the outlet of Glacial Lake Winooski 50 to 60 years before the ice retreated north of Muzzy Brook at North American Varve 6658 putting the beginning of Glacial Lake Winooski at approximately North American Varve 6600 which corresponds to about 14,220 yr BP (Fig. 14). Therefore , Glacial Lake Winooski lasted ~375 years, expanding both north and west of its outlet before the ice dam in the Winooski

	Calculated Ice Margin Retreat Rates	Distance (m)	Time (years)	Ice Retreat Rate (m/year)
A	Muzzy Brook to Montpelier	5,900	20	295
B	Montpelier to Wrightsville Reservoir	6,400	21	305
C	Upper Connecticut River Valley (Ridge et al., 2012)			300
D	Estimated time for ice margin to retreat from Glacial Lake Winooski Outlet to Muzzy Brook if retreat rate was 300 m/year.	15,000	50	300
E	Estimated time for ice margin to retreat from Glacial Lake Winooski Outlet to Muzzy Brook if retreat rate was 250 m/year.	15,000	60	250

Figure 15: Calculated ice sheet retreat rates between (A) Muzzy Brook and Montpelier and (B) Montpelier and the Wrightsville Reservoir based on the time interval between the initiation of sedimentation at each site and the distance between these sites (Fig. 14). (C) Average ice sheet retreat rate in the upper Connecticut River valley (Ridge et al., 2012). (D, E) Calculated time necessary for the ice sheet to retreat from the outlet of Glacial Lake Winooski (the Williamstown Gulf threshold) to a position in the Second Branch equivalent to the Muzzy Brook section based on a retreat rate of 300 m/yr (D) and 250 m/yr (E).

River valley retreated far enough west to uncover the Gillette Pond outlet allowing water levels to rapidly fall and the Williamstown Gulf outlet to be abandoned (Figs. 6-9, 14). These dates are consistent with Ridge's (2016) 14.1 and 13.9 ka timelines across the Winooski drainage basin (Fig. 12).

Further confirmation of the timing of ice retreat across the region comes from exposure age dating. In a study focused on sampling an elevation transect from Mount Mansfield in the northern Green Mountains (western flank of Glacial Lake Winooski, Fig. 12), Corbett and others (2019) found that 15 samples collected between ~400 and 1,200 m a.s.l. have indistinguishable ^{10}Be ages averaging 13.9±0.6 ka indicating rapid thinning of the ice sheet at this time. This date range of rapid ice sheet thinning coincides with the ~14,170 to 13,800 ka life span of Glacial Lake Winooski when the growing lake tracked the rapidly retreating ice sheet east of the mountains. Exposure ages from elevation transects across New England similarly show 800-1,000 m of ice thinning over time spans too short to distinguish with this dating technique (Halsted et al., 2022).

The Littleton-Bethlehem (Middlesex?) Readvance in the Winooski River Basin:

The Littleton-Bethlehem Readvance of the ice sheet is well documented in northern New Hampshire by a widespread array of moraines (the White Mountain Moraine System) and a stratigraphy of till with thrust-faulted sand and gravel within Glacial Lake Hitchcock sediments near the Comerford Dam in the upper Connecticut River valley (Thompson et al., 2017; Thompson et al., 1999). Thompson and others (2017) associate this readvance with a 123 year long interval in the North American Varve Chronology (14,021-13,898 cal yr BP) that corresponds with the Older Dryas cooling event. This same readvance is also referred to as the Middlesex Readvance and appears as the 13.9 ka timeline in Figure 12 (Ridge, 2016; Ridge et al., 2012).

In the Winooski River basin many sites contain one or more layers of diamict interlayered with lake sediments and these have been cited as evidence of ice sheet readvance corresponding to the Littleton-Bethlehem (Middlesex) readvance (Dunn and Springston, 2019; Larsen et al., 2003a; Wright, 2015a). However, in many of these sections there's no stratigraphic evidence recording ice advance or retreat below or above these diamict layers and landslides from the surrounding steep hillsides may be a more likely explanation for the diamict layers (Wright, 2023). Of particular note is a diamict overlying intensely sheared Glacial Lake Winooski sediments along Culver Brook adjacent to the Wrightsville Reservoir (Fig. 16; Larsen, 1999a; Stop 5C). Larsen (2001) dated a piece of wood found in the overlying till at 11,900 ±50 C-14 yr BP which corresponds to a calibrated age range of 13,675-13,791 yr BP, well after Glacial Lake Winooski had drained (Fig. 8). For this diamict to be a readvance till the ice sheet needed to advance and subsequently retreat across almost the entire Winooski River basin. While doing so it would have dammed the basin a second time, yet there's no evidence of a younger widespread "Glacial Lake Winooski." It seems more likely that the upper till at Culver Brook is a landslide deposit that occurred well after the ice had retreated from the area and climate had warmed sufficiently for the surrounding hill slopes to be tree-covered.

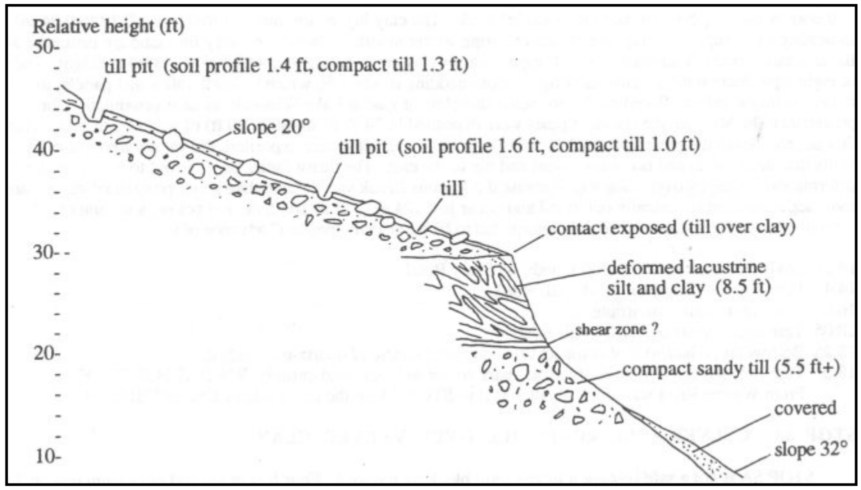


Figure 16: Larsen's (1999) cross-section of the Culver Brook 2-till site. The dated wood was found in the upper till a short distance above the deformed lacustrine sediments. View looks northwest.

Our mapping elsewhere in northern Vermont indicates that landslides were common following ice retreat (Wright, 2023). A recent landslide (May, 2019) along Cotton Brook on the eastern slopes of the Green Mountains (~6.2 km SE of Stowe Village) reveals a stratigraphy consisting of multiple landslide deposits interbedded with undeformed lake sediments (Fig. 17). Some of the landslide deposits are diamict and others consist of faulted blocks of layered sand/silt that slumped into the lake basin. Note that the top several meters consists of diamict (colluvium) and before the landslide revealed the underlying lacustrine sediments, Wright (2018b) mapped this entire hill slope as till. Both recently deposited, unvegetated till and thick accumulations of lacustrine sediments are unstable on steep mountain slopes, particularly when water saturated. Whether triggered by storms, spring thaw, high groundwater pressure, or some other process, landslide deposits are common both within the

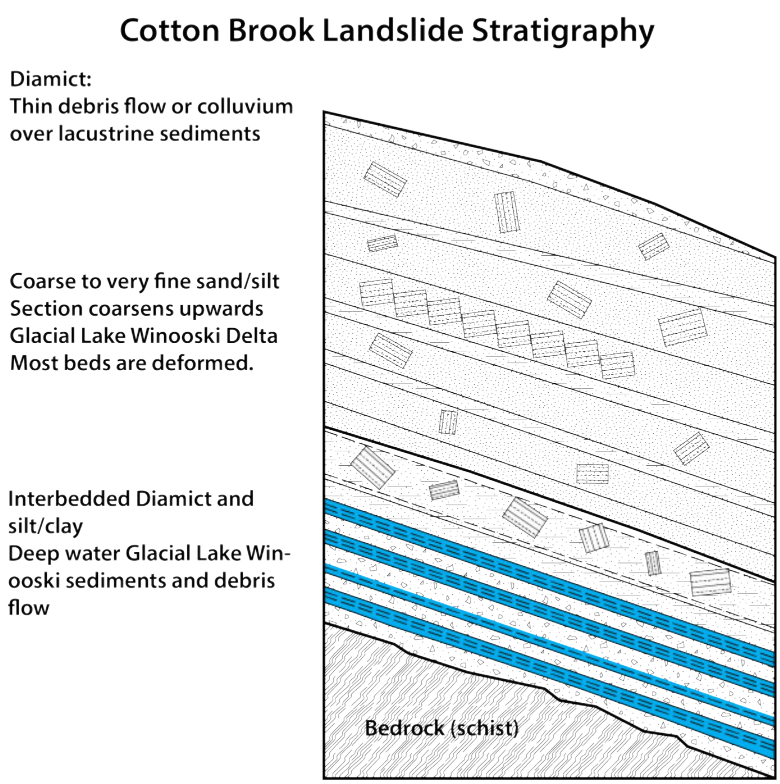


Figure 17: Diagrammatic sketch of Glacial Lake Winooski sediments deposited in the Cotton Brook valley. Multiple landslide deposits within the lake basin are separated by undeformed lake sediments.

region's glacial lakes and on the steep slopes adjacent to these lakes. That said, many if not most of these deposits consist of diamict and distinguishing an in-place till vs remobilized till (diamict) is difficult or impossible in most areas. Detailed structural work designed to ascertain the direction of shear strain in deformed sediments underlying the diamict layers could help distinguish diamicts deposited by readvancing ice from those deposited by landslides. This work has not yet been attempted.

Ice-Contact Environment During Deglaciation

As the ice sheet thinned ice flow was directed parallel to the region's valleys, the mountains emerged first as isolated nunataks and then as continuous ridges. We have mapped a variety of landform/sediment associations that likely formed along or close to the ice margin and provide some insight into processes occurring there.

Moraines

In northern Vermont traditional recessional moraines, single or intersecting ridges largely composed of till, occur only rarely across valley bottoms and are small in scale. The near absence of this type of moraine is a likely consequence of the rapid ice sheet retreat rate across northern Vermont (~300 m/yr, see previous section). Prominent moraines, similar to White Mountain Moraine System associated with the Older Dryas readvance (Thompson et al., 2017; Thompson et al., 1999), have not yet been identified in the region.

While traditional moraines with positive relief are rare, another type of ice marginal landform is widely developed on many mountain slopes in the Green Mountains and may constitute a different style of moraine (Wright, 2019b). These are asymmetric steps or benches, the hill slope sequentially steepens and flattens, consisting almost entirely of till (diamict) without any positive relief (Figs. 18, 19). These landforms generally contour across the valley sides

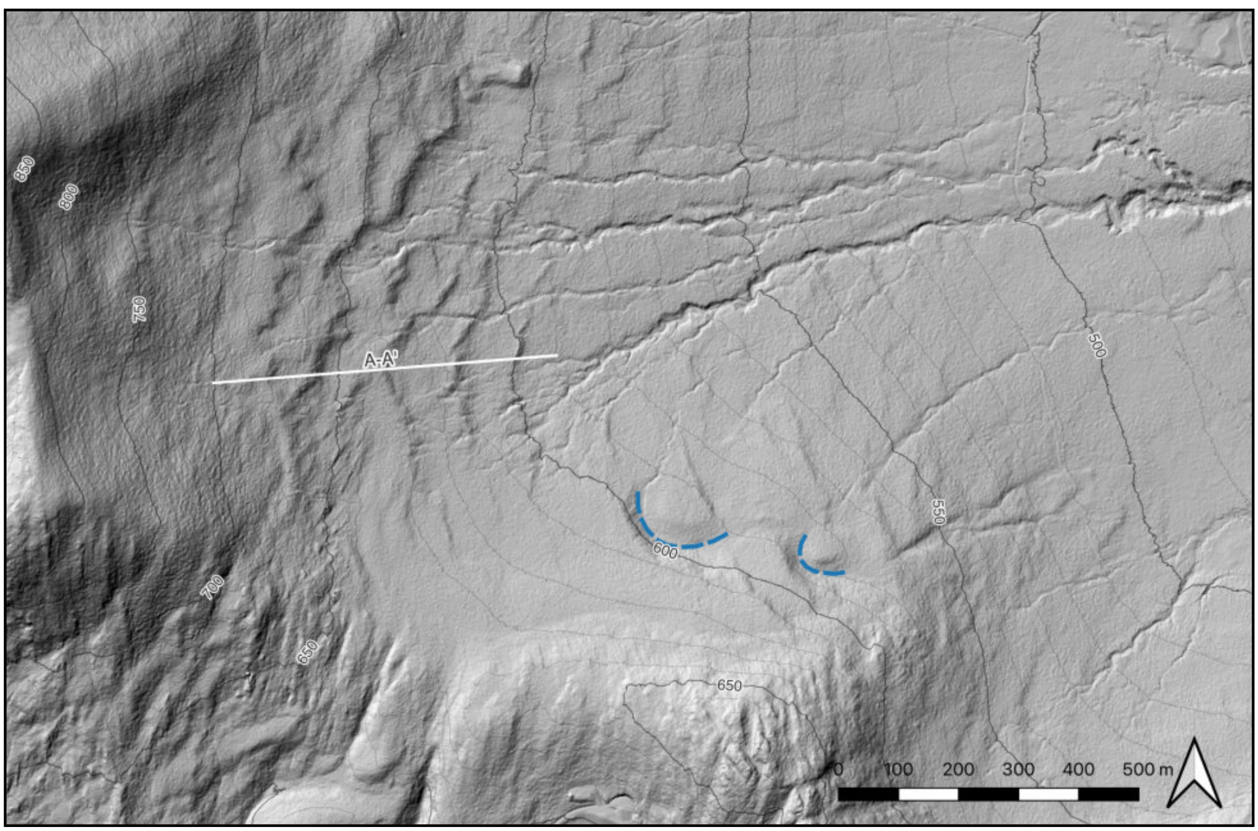


Figure 18: Northwest lighting shadows the steep faces of nested till benches in the Sterling Brook valley, east flank of the Green Mountains. These landforms are most prominent where till cover is thick and slopes are steep. Farther down slope the magnitude of individual steps lessens and the landform eventually disappears. Blue dashed lines highlight meltwater channels. White line marks topographic profile A-A' (Fig. 19).

Figure 19: West-East topographic profile across a flight of stepped till benches in the Sterling Brook valley that are interpreted to be an ice-marginal landform (a type of moraine?) that formed perhaps annually as the ice sheet thinned. Note that treads of each step are still gently sloped and the elevation change between adjacent treads ranges from ~9-12 m. See map for location of profile (Fig. 18). No vertical exaggeration.

and commonly occur as flights of subparallel steps in areas where the till cover is thick enough to bury the underlying bedrock and on slopes range from ~12 to 20% (Figs. 18, 19). Steeper slopes are generally dominated by bedrock. These landforms are not well developed on gentle slopes <10% (Fig. 18). In some areas these landforms are cut by meltwater channels. Field work designed to understand the internal structure of these landforms has generally been unsuccessful. However, thin beds of silt/very fine sand occur in small exposures within the diamict, perhaps deposited in ice-marginal ponds, and these beds are universally deformed and generally dip gently upslope.

Wright (2019b, 2022d) proposed that these steps form when till, recently exposed as the ice sheet thins during the summer months, (1) flows down-slope and accumulates at the ice sheet margin and/or (2) is pushed into a ridge as the ice sheet expands upslope during the winter months. The thin layers of silt/fine sand may be deposited in small pools of meltwater collecting at the ice sheet margin. The upslope dip of these layers argues for (1) a limited degree of push from expanding ice or (2) that deformation may result from water-saturated till being squeezed from beneath the ice sheet margin. Till extrusion to the ice sheet margin and crevasses may also be a significant process and may explain some of the irregular patterns displayed by landforms. Davis and others (2022) refer to similar landforms they mapped on Mount Moosilauke as “till benches” and emphasize the role of till slumping against the ice sheet margin as key to their formation (Davis, 2024). While Geier (2021) suggested that these benches were produced by solifluction, Davis (2024) argues that these benches lack the distinctive pattern of solifluction lobes. We have similarly not observed any benches in the Green Mountains displaying features characteristic of solifluction lobes.

Where traditional recessional moraines record periods of time when till has accumulated at the ice margin, e.g. during a standstill or an advance, these benches/moraines, where well preserved, may record yearly thinning of the ice sheet. Along the flanks of Mount Mansfield and Belvidere Mountain Wright (2019b) measured vertical drops of 9-16 m between adjacent benches/moraines. Near the Québec border flights of meltwater channels drop 3 to 10 m between adjacent channels and may also record yearly ice sheet thinning (Wright, 2024). If these elevation drops are a measure of yearly ice sheet thinning, then 1,000 m of the adjacent mountains could have been exposed in just a few hundred years. While that rate seems very fast, it may be consistent with the calculated average ice retreat rates of ~300 m/yr and explain why exposure age dates show little change with elevation across New England (Corbett et al., 2019; Halsted et al., 2022).

Ice-Marginal Channels, Kames, and Ice-Contact Deltas

Rapid thinning and retreat of the ice sheet across northern New England generated tremendous volumes of meltwater that flowed both beneath and along the margins of the ice sheet. Ice-marginal channels document the pathways this meltwater followed. A well-preserved set of meltwater channels occurs in the southwest corner of the Waterbury Quadrangle (Fig. 21; Wright et al., 2023a). These south- and west-sloping channels are parallel to the inferred margin of the ice sheet. These channels occur in sub-parallel sets and were likely sequentially eroded into recently exposed till by streams flowing parallel to the edge of the ice sheet as the ice sheet thinned from year to year (Fig. 20).

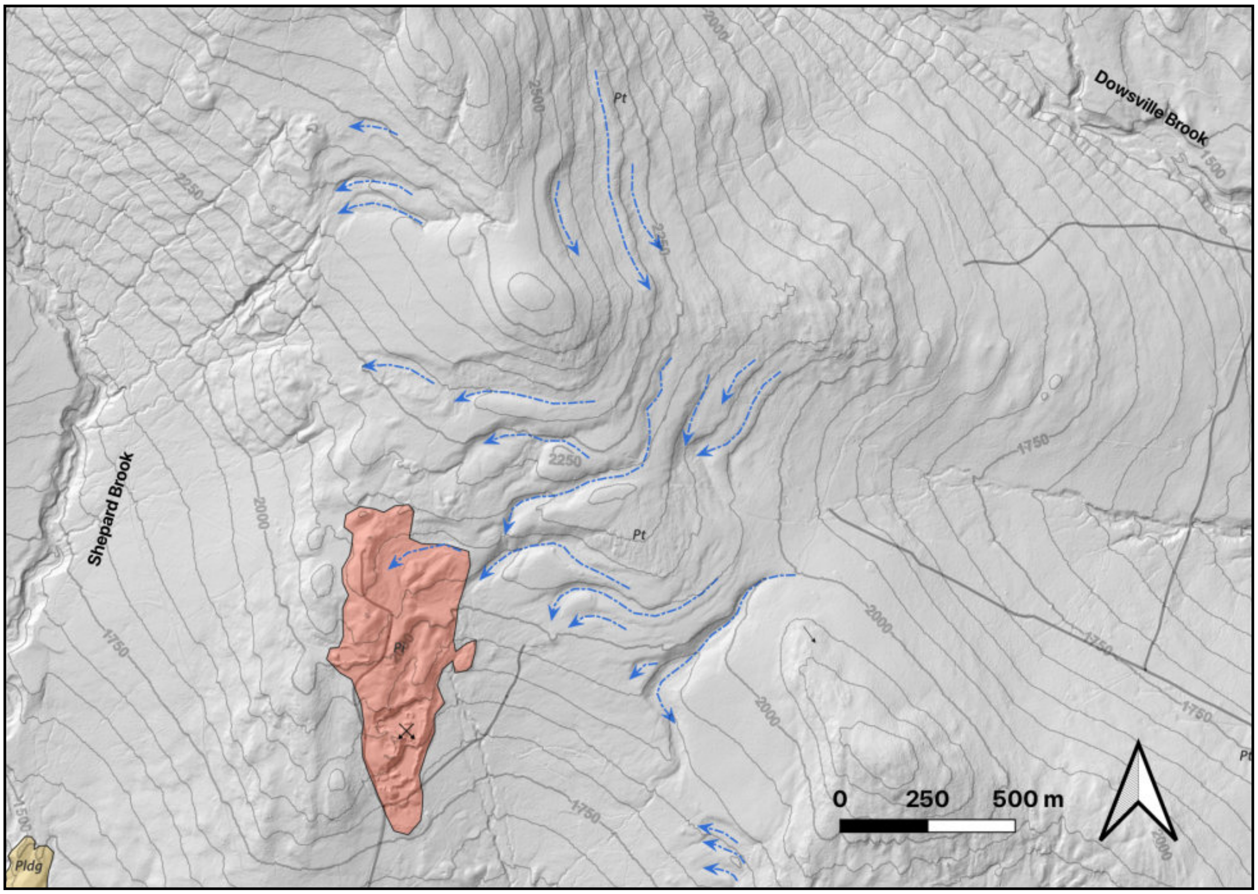


Figure 20: Dashed blue arrows mark nested meltwater channels developed on the eastern flank of the Green Mountains in the southwest corner of the Waterbury Quadrangle. These channels record meltwater flowing along the margin of the ice sheet as it thinned from year to year. A pit is developed in a deposit of ice-contact sand and gravel (Pi). A small Glacial Lake Granville delta (Pldg) is also shown in the Shepard Brook valley. Contours in feet.

In addition to eroding channels, that same meltwater transported large volumes of sediment. Most of this sediment was eroded from till occurring beneath, on the surface of, or along the edge of the ice sheet, but also included debris entrained within the ice that was released as friction from moving water melted the adjacent ice. Kames, deposited by streams flowing between the edge of the ice sheet and adjacent mountain slopes, are particularly well developed along the western slopes of the Green Mountains, i.e. the eastern side of the Champlain valley, perhaps due to the large scale of the glacier in that valley (Connally and Calkin, 1972). The western extent of our recent mapping just reaches into the Champlain valley and this is where these landforms are best developed, albeit on a much smaller scale than at lower elevations farther west (Fig. 21). During times when the ice sheet bordered and/or dammed a glacial lake, sediments carried by meltwater streams formed ice-contact deltas where they entered that lake (Fig. 21). Several of these deltas were mapped where ice-marginal streams entered relatively small, short-lived glacial lakes dammed by the retreating ice along the western slopes of the Green Mountains (Wright, 2022c, d). Distributary channels are often preserved on the delta surfaces. Kettles are also common in both kames and ice-contact deltas (Fig. 21).

Eskers

Eskers are common in the region and generally follow the major stream valleys either NW-SE across the mountains or N-S parallel to the mountains. This is consistent with the directions of ice sheet flow and subsequent hydraulic gradients as the ice sheet thinned and its flow directions were dictated by the region's valleys.

Ice flow across the mountains is highlighted by he Miller Brook esker which will be visited on this field trip (Stop B4). Meltwater flowing through this esker tunnel was sourced from ice in the Champlain valley, routed through Nebraska Notch, a low point in the Green

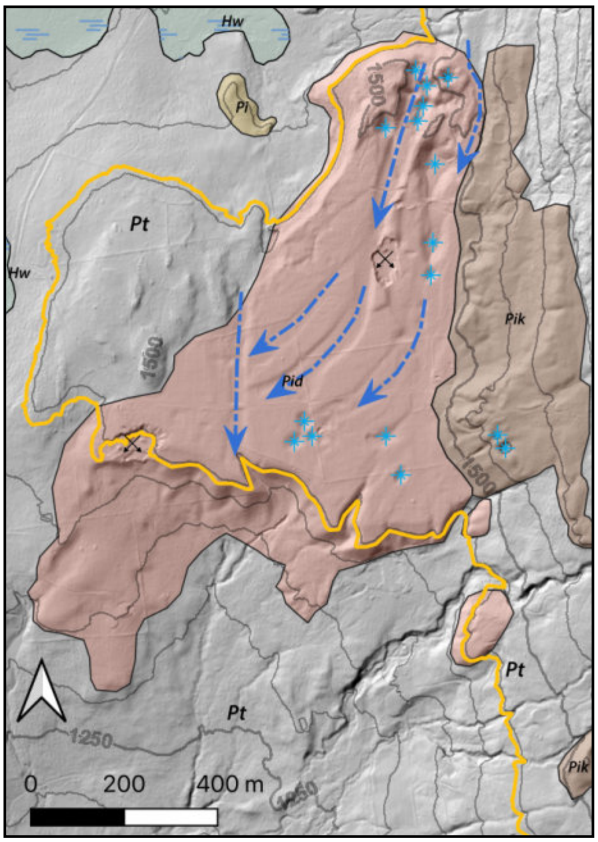


Figure 21: Numerous kettles and distributary channels (dashed blue lines) occur in this ice-contact delta (Pid) ~2.5 km east of Lincoln, Vermont near the Battle trailhead. This delta was deposited in a relatively small glacial lake (outlined with the orange line) with an outlet at Alder Brook. Delta is bordered by sand and gravel comprising a kame (Pik) deposited along the margin of the ice sheet shortly before the delta was deposited. Contours in feet.

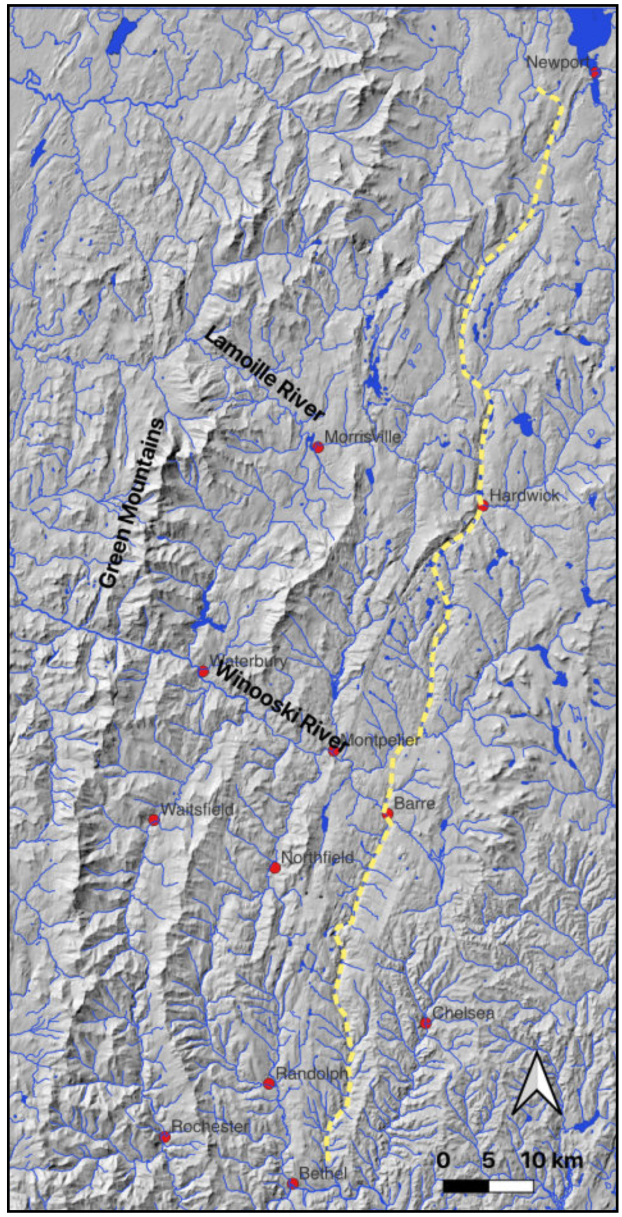


Figure 22: A long esker (dashed yellow line), discontinuously exposed along Vermont Route 14, is evidence of a long-lived, time-transgressive subglacial drainage system active during ice sheet retreat.

Mountain ridge line, and then down the Miller Brook valley.

The Winooski River valley, prominently cutting WNW-ESE across the mountains, also hosted an esker. While a distinct esker landform no longer exists, its existence is indicated by numerous exposures of subaqueous fan deposits sourced from this esker, sedimentary structures preserved in a gravel pit in Middlesex, and numerous water well logs in the valley that bottom in coarse sand and gravel, i.e. the esker is buried beneath younger lacustrine sediments.

One of the best examples of an esker oriented N-S is a long segmented esker that follows Vermont Route 14 completely across the Montpelier Sheet and probably to the Québec border. This esker threads its way along numerous aligned stream valleys and up and over many low-elevation drainage divides (Fig. 22). While many portions of this esker have been quarried away, significant portions still exist and will be visited during the Friday field trip. The significance of this esker is that it demonstrates the evolution of a long-lasting, time-transgressive subglacial drainage conduit within the ice sheet during ice retreat. Given that eskers are time-transgressive depositional landforms, i.e. at any one time net accumulation of sediments likely occurred only in the last several kilometers of the subglacial ice tunnel, the up-glacier tunnel system itself must have continued to migrate northward as the locus of deposition in the subglacial tunnel also shifted to the north. In areas where no esker was mapped, it's unclear whether (1) an esker was never deposited at these locations, (2) an esker was deposited but was later eroded away or fully excavated by human activities, or (3) the esker lies buried beneath younger, lacustrine deposits.

Alluvial Fans

Alluvial fans occur on a wide-range of scales across northern Vermont. These form where sediments eroded from the valley sides, generally glacial till, have been carried downhill by tributary streams and deposited where the stream gradient abruptly lessens where it flows onto the valley bottom. The apex of these fans frequently consists of debris flow deposits. Farther down-fan sediments consist of lenses of sand/gravel that may fine to silt at their toes. In areas above the elevation of glacial lakes, the fans are deposited on till. More commonly they were deposited on lake bottom sediments and both concurrent with the lakes and the much longer period of time following. In many places these fans are large enough to extend almost completely across the valley and the courses of the local streams flow around the toes of these fans. Work on alluvial fans in northern Vermont suggests that fans have been episodically active throughout the Holocene and many received their most recent pulse of sediment following European land clearing in the late 18th and early 19th centuries (Bierman et al., 1997; Jennings et al., 2003). Related work by Noren et al. (2002) recording pulses of clastic sediment deposited in ponds and small lakes, indicates that pre-European settlement erosion has not been uniformly distributed throughout the Holocene and seems instead to be concentrated during periods of increased high-intensity storms. If climate shifts produce a greater frequency of high-intensity storms, further sedimentation on the area's alluvial fans seems likely

From Till to Forest: The Evolution of Climate and Vegetation in Central Vermont

Laurie D. Grigg

Sediment cores from upland lakes and wetlands in central Vermont preserve an archive of post-glacial environments and provide insights on past changes in climate and vegetation. This presentation focuses on the transition from the late-Pleistocene to early Holocene in three records from the region, Twin Ponds, Brookfield, VT (Grigg et al., 2021; Mandl et al., 2016), Pecks Pond, Barre, VT (Magdon et al., 2018), and Knob Hill Pond, Marshfield, VT (Oswald and Foster, 2011; Oswald et al., 2018). The oldest radiocarbon ages from these records range from 13.0 to 13.7 cal ka BP but sediment recovery extends another 40-80cm suggesting these are minimum ages. Rapid warming centered around 11.5 cal ka BP, as well as, pre- and post-Younger Dryas centennial-scale climatic variability are prominent in all records. Multi-proxy records from Pecks Pond indicate an early warming (>13.3 cal ka BP) evident by an increase in $CaCO_3$, aquatic vegetation, and bivalves. These results suggest that vegetation flourished in aquatic environments prior to terrestrial environments. Several centennial-scale, lake-level fluctuations at Pecks Pond are also evident >13.3 cal ka BP and may indicate variable precipitation that may have impacted the glacial mass balance of the retreating Laurentide Ice Sheet. At the start of the Younger Dryas cooling event (YD; ca. 12.9 cal ka BP), Pecks Pond fluctuated from shallow lake to wetland and back to shallow lake. During the YD, all proxies from Twin Ponds show evidence for generally cold but not necessarily reversing conditions and increased meltwater runoff. The Twin Ponds ostracod record shows a northward shift in the position of the summer polar front during the latter half of the YD which is consistent with evidence for increasing sea-ice instability in the northwestern North Atlantic. There is limited evidence for additional cool intervals centered around 10.6, 10.2 and 9.6 cal ka BP at Twin Ponds that may reflect continued instability in the North Atlantic during the final phases of deglaciation.

Pollen records from Twin Ponds (Grigg et al., 2021) and Knob Hill Pond (Oswald and Foster, 2011) suggest that central Vermont did not experience a significant tundra phase. Very low pollen accumulation rates prior to the YD suggest limited sedge, grass, and spruce or the long-distance transport of pollen grains from southern regions. Sedge pollen at the base of the pollen record from Twin Ponds indicates the establishment of a fringing wetland around the pond or limited patches of sedge tundra. During the YD, pollen accumulation rates of spruce increased slightly but herbaceous taxa remained low suggesting open or isolated patches of spruce amongst a largely till-dominated landscape. It was not until 11.5 cal ka BP at the end of the YD that pollen accumulation rates increase to levels comparable to a closed forest. This transition includes a sequences of non-analogue mixed forest types that culminated in the pine-oak forest characteristic of the early Holocene (Grigg et al., 2023).

Guide to Field Trip Stops

Friday Afternoon Field Trip June 7, 2024: Second Branch of the White River: East Randolph to Williamstown

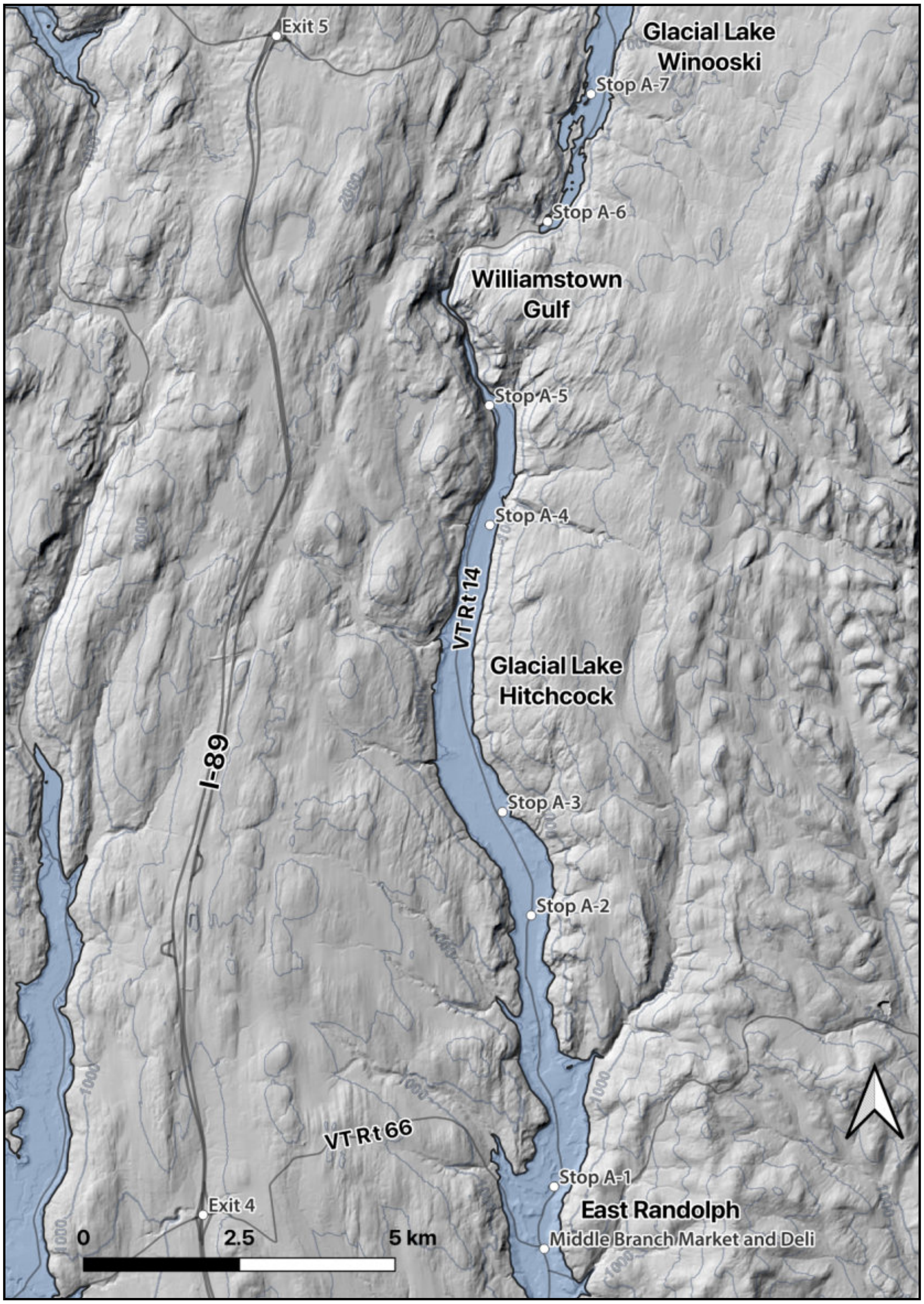


Figure 23: Stop Map for Friday field trip is centered on the Second Branch of the White River.

Introduction

This field trip follows the 2nd Branch of the White River to its source in Williamstown Gulf and then continues a short distance down the Stevens Branch of the Winooski River (Fig. 23). The White River valley hosted an arm of Glacial Lake Hitchcock whereas the Stevens Branch valley an arm of Glacial Lake Winooski and the drainage divide between them was the outlet of Glacial Lake Winooski. Integral to the glacial history of these valleys is a subglacial drainage system manifest by an esker and a variety of ice-proximal lacustrine sediments sourced from that subglacial tunnel. The objective of this field trip is understand the intertwined histories of the two lakes and the esker system. The interpretations presented here are derived from mapping in the Brookfield Quadrangle (Wright, 2022a, b).

Stop A-1: Ice-Proximal Glacial Lake Hitchcock sediments, East Randolph

This gravel pit displays a beautiful section of ice-proximal sand and gravel deposited in Glacial Lake Hitchcock. Immediately north of this pit is a second abandoned pit that formerly exposed the top of an esker (Fig. 24). Bedding initially drapes over the esker, but quickly becomes generally horizontal as the esker was rapidly buried. The sediments in the active pit consist of interlayered medium to coarse sand and gravel most likely deposited close to the ice margin where the sediment size varied with discharge from the mouth of the tunnel hosting the esker.

Figure 24: Crest of esker in East Randolph, Vermont is overlain by coarse ice-proximal sediments deposited in Glacial Lake Hitchcock. View looks south.

Stop A-2: Crest of Route 14 Esker

Vermont Route 14 generally follows the course of the esker up the valley. At this road pullout we're parked on the crest of the esker that rises above bordering alluvium and wetlands sediments. This esker is part of a very long esker follows the route of Vermont Route 14 between Randolph to the south and Coventry to the north and may well extend farther in both directions (Fig. 23). Here, as in many parts of New England, the esker provided the logical, well-drained pathway for the road to follow along the valley. An abandoned gravel pit (now a junk yard) lies a short distance farther north on the west side of the road. Inspection of 1960's era stereo aerial photography revealed an esker formerly existed here, but has been completely quarried away.

Stop A-3: Ice-Proximal Lacustrine sediment in small pit
This small gravel pit is actively being converted to a new house. Interlayered coarse pebble gravel and medium to fine sand is still exposed lying in direct contact with the underlying bedrock. A larger outcrop with overlying till is no longer visible.

Stop A-4: Mid-Valley Gravel Pit
This pit lies in an elliptical mound in the middle of the valley. The gravel here consists of coarse sand, pebble, cobbles, and small boulders, sediments far too large to have been deposited by the modern river which is a quite small stream. Exposures in a manure pit on the farm immediately to the south show that the gravel lies unconformably on silt/clay, quiet water Glacial Lake Hitchcock sediments. These sediments must have been deposited as a large-scale bar by a high-discharge stream sometime after Glacial Lake Hitchcock drained from the valley. The likely source of that water is drainage from Glacial Lake Winooski which collected both subaerial drainage from both the Winooski and Lamoille River basins and ice sheet meltwater during it's ~375 year long history. The gravel in this pit may be sourced from a delta that formed where Glacial Lake Winooski drainage entered Glacial Lake Hitchcock a short distance north of this pit. When Glacial Lake Hitchcock drained from the valley, drainage from Glacial Lake Winooski eroded the previously deposited delta and redeposited those sediments in across the former bottom of Glacial Lake Winooski.

Stop A-5: Base of Williamstown Gulf
Upstream from this stop the valley dramatically narrows and steepens as it follows a course likely controlled by bedrock structures in the underlying Devonian Waits River Formation. Hidden in the trees on the opposite (north) side of the valley is a terrace composed of sand and gravel at an elevation of 254 m and a similar, albeit narrower, terrace lies immediately south of this stop at an elevation of 253 m. Both of these terraces are only slightly higher than the projected elevation of Glacial Lake Hitchcock (Koteff and Larsen, 1989). Unless these are kames that formed along the ice margin, they must have formed when the valley was filled with sediments to this elevation, most likely a delta in narrow, shallow arm of Glacial Lake Hitchcock. Sediments deposited in this delta were likely sourced predominantly from the same subglacial tunnel that deposited the esker farther south in the valley. Glacial Lake Winooski outflow across the bedrock threshold probably did not entrain much sediment, but that same water could have eroded previously deposited till along the sides of Williamstown Gulf between the threshold and this stop.

Stop A-6: Glacial Lake Winooski Outlet
This is the drainage divide (276 m, 906 ft) between the Second Branch of the White River flowing south from here and eventually entering Long Island Sound and the Stevens Branch of the Winooski River flowing north from here and eventually flowing into the Gulf of St Lawrence. This is the lowest drainage divide in the Winooski River drainage basin and was the outlet of Glacial Lake Winooski. The topographic drainage divide is actually a few hundred meters north of here, and lies on the toes of two converging Holocene alluvial fans that did not exist when Glacial Lake Winooski existed. As noted earlier, water flowing across this outlet was sourced both from meteoric water entering both the Winooski and Lamoille drainage basins and the rapidly melting ice sheet which was retreating at ~300 m/yr. One challenge that exists when projecting the former lake surface from this divide across the drainage basin is to estimate the depth of water flowing across this outlet. This depth varies with all lakes both seasonally and with storms. Given the sources of water noted above, a water depth estimate of at least 5 m across this threshold seems reasonable, but we have not made any quantitative estimates.

Stop A-7: Route 14 Esker, Williamstown Transfer Station
This last stop is at the south end of a very large disused gravel pit. The source of the sand and gravel was the same esker visited farther south earlier on this trip and the ice-proximal subaqueous fan deposits that blanketed the esker. An excavation on the west side of the esker reveals some of its internal structures. Large-scale, south-dipping cross-beds indicate south directed water flow in the esker tunnel, parallel to the local slope of the ice sheet surface. Faults within the sediments may be due to a loss of support as ice beneath or adjacent to sediments melted. Farther north in the pit sand and gravel are deposited directly on water-eroded bedrock, evidence that the subglacial stream effectively eroded all of the till, one of the primary sources of sediment deposited by that stream both in the subglacial tunnel and in the lake.

Saturday June 8, 2024: Stowe/Middlesex Geology

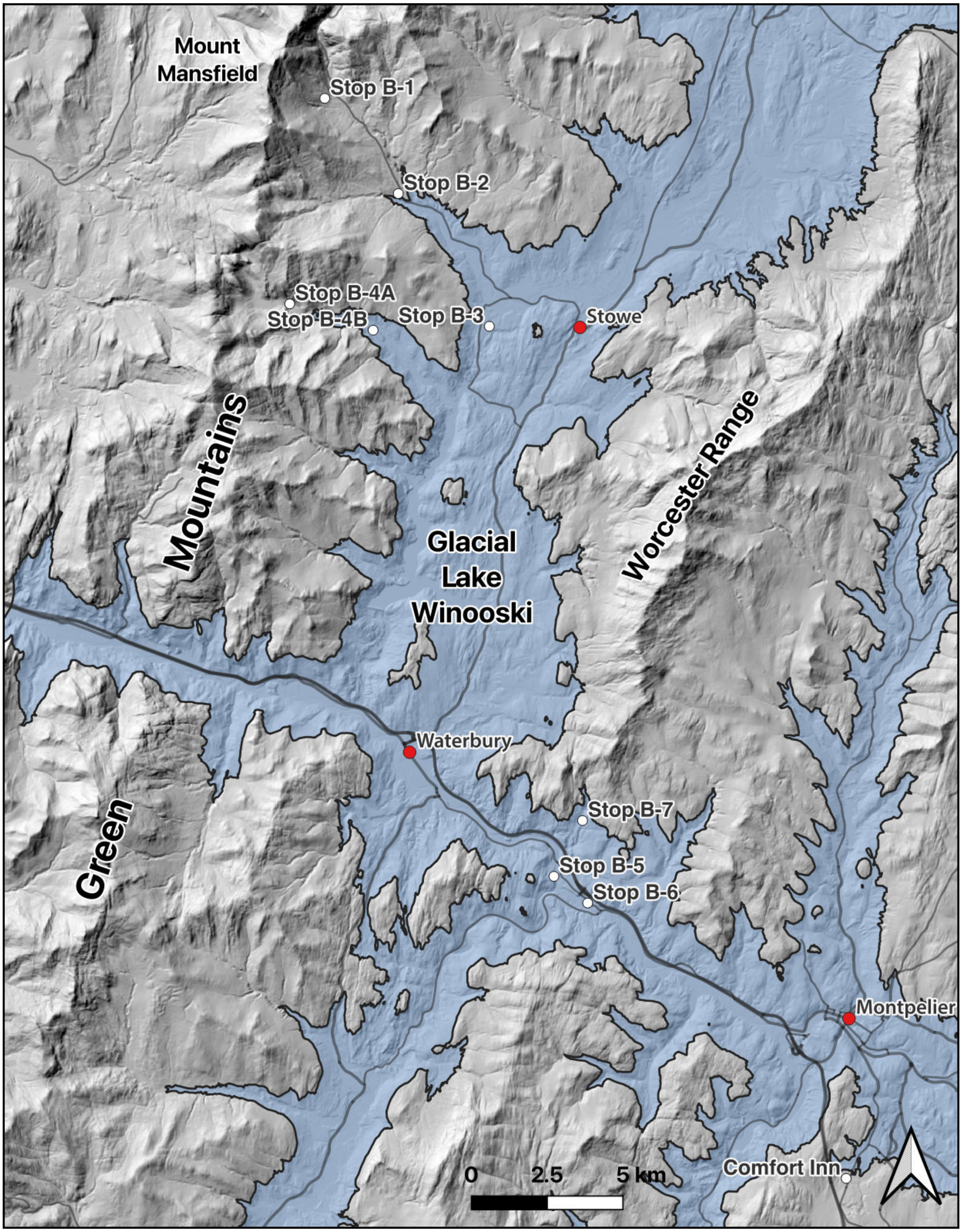


Figure 25: Map shows the locations of Saturday field trip stops.

Stop B1: Till Benches/Moraines on Mount Mansfield: Midway Lodge Parking Lot, Stowe Mountain Resort

One of the striking landforms visible in the LiDAR imagery are subparallel benches of till that contour across many mountain sides in the area (see earlier discussion). At the Stowe Mountain Resort the northernmost downhill ski trail, Chin Clip Runout, fortuitously cuts across a flight of these landforms (Figs. 26). Cross-sections of half a dozen till benches are visible while walking up the ski trail from the Midway Lodge parking lot, near the base of the Gondola lift. A small stream cut in the third bench up from the base exposes a thin layer of very fine sand/silt within the till. A topographic profile highlights the stair-step profile of these landforms as the slope alternates between gentle and steep (Fig. 27). Along this particular profile the average distance between these 6 adjacent benches is ~75 m and the average elevation drop is ~16.6 m.

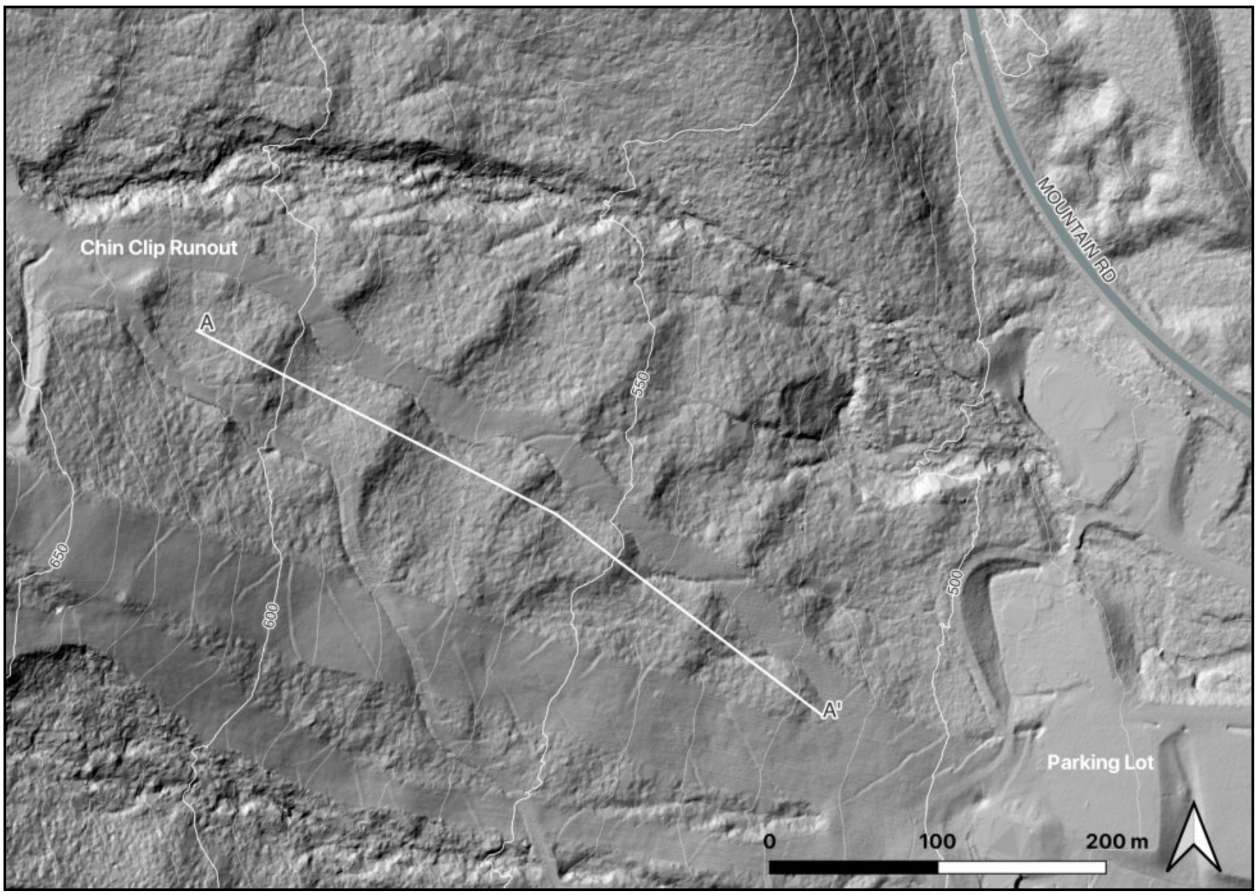


Figure 26: Map shows the Chin Clip Runout ski trail cutting across a set of subparallel till benches near the base of the Stowe Mountain Resort gondola. Topographic profile A-A' is appears on following page. Contours are in meters.

These landforms lack the lobate outline of solifluction lobes. Their occurrence in subparallel sets stepping down the hillside suggests that they formed as ice-marginal landforms as the ice sheet thinned. The geometry of the landform, layers of silt within the diamict, and folded diamict layers suggest that the diamict within these landforms was transported after it was initially deposited by some combination of (1) slumping from upslope or the ice sheet surface and accumulating along the ice margin, (2) extruding from beneath the ice sheet margin, and (3) being pushed by the ice sheet as it expanded during the winter months. As noted earlier, exposure age dating on Mount Mansfield Corbett and others (2019) found that 15 samples collected between ~400 and 1,200 m a.s.l. have

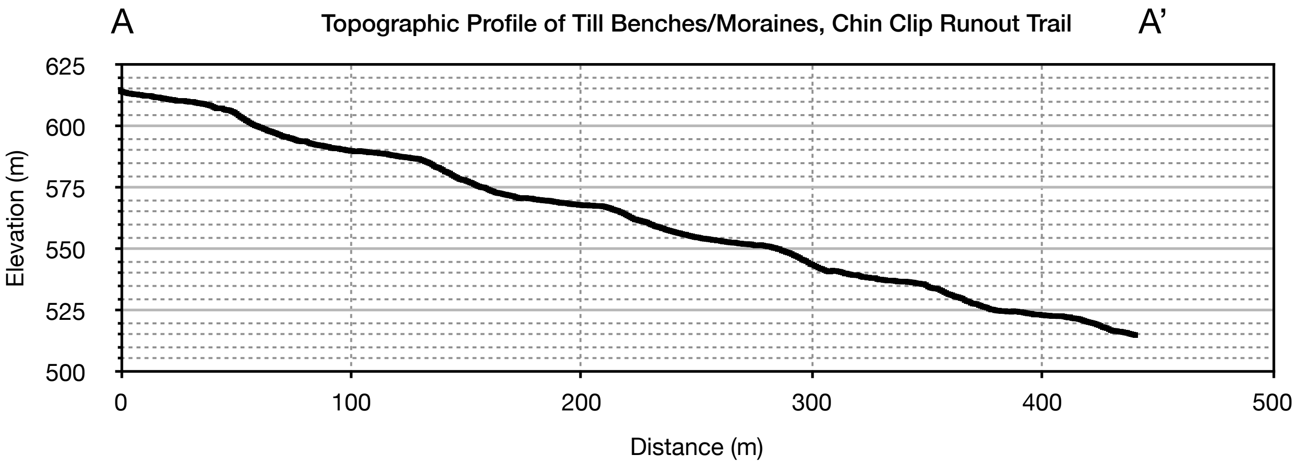


Figure 27: Topographic profile of the till benches visible along the Chin Clip Runout ski trail. Vertical exaggeration ~1.25X.

indistinguishable ^{10}Be ages averaging 13.9±0.6 ka indicating rapid thinning of the ice sheet when these till benches formed.

Stop B-2: Glacial Lake Winooski Delta: Stowe Nordic Center Parking Lot

As we travel back down the Mountain Road (Vermont Route 108) we will cross a Glacial Lake Winooski delta that underlies the Stowe Nordic Center and general employee parking lot. This terrace is underlain by a thick section of sand and gravel and slopes gently south. The elevation of the terrace ranges from ~342-347 m (1,122-1,138 ft), very close to the projected elevation of Glacial Lake Winooski. The projected lake shoreline here has been isostatically tilted upwards ~65 m (~213 ft) from the elevation of the lake outlet in Williamstown Gulf (Figs. 6-8).

While small, short-lived lakes were likely trapped in many of the surrounding mountain valleys, Glacial Lake Winooski was the highest elevation wide-spread lake to flood the Winooski River basin including the valley lying between the Green Mountains and the Worcester Range. Glacial Lake Winooski first formed ~14,220 years ago, lasted ~375 years, and lowered to the elevation of Glacial Lake Mansfield 1 ~13,850 years ago (Figs. 14, 15). The time span of the lake is centered on the 13,900 year exposure age dates from Mount Mansfield (Corbett et al., 2019). During this period of time the lake was rapidly expanding to the north following an ice margin retreating at ~300 m/yr.

Stop B-3: Glacial Lake Mansfield Delta/Buried Channel: Stowe High School

The Stowe High School is built on a terrace that we interpret as a Glacial Lake Mansfield delta (Wright, 2020). The implication is that the West Branch of the Little River valley which we just drove down was largely filled with ice contact and lacustrine sediments when Glacial Lake Winooski lowered to the elevation of Glacial Lake Mansfield 1. The West Branch of the Little River then deposited alluvium on the former lake floor which graded to a delta here where the river flowed into the lake.

The sediments here bury the old bedrock channel of the Little River which lies under the school and is revealed by both the school's water well and a recent Tromino passive seismic survey. The modern Little river flows east from here until it reaches the Village of Stowe where it turns south cutting a new bedrock floored channel. The waterfall in the village is the reason Stowe developed where it did.

Stop: B4 (A-I): Miller Brook Esker

The Miller Brook Valley, Stowe, Vermont, is a generally ESE-draining valley extending from a low point along the crest of the Green Mountains (Nebraska Notch, Elev. 580 m, 1900 ft) to its confluence with the Little River (Elev. 184 m, 605 ft), (Fig. 25). The objective of this field stop is to show participants the key landforms and sediments utilized to interpret the glacial history of this valley. These include large-scale cirque-like landforms visible at the head of the valley, smaller-scale erosional and depositional landforms indicative of a major subglacial drainage system, and subaqueous fan and more ice-distal sediments deposited in an arm of Glacial Lake Winooski. Portions of the Miller Brook valley have been visited on previous NEIGC field trips (Connally, 1972; Dunn et al., 2011; Wagner, 1972; Wright et al., 1997). The following stop description is a shortened version of a longer NEIGC field guide to the valley (Wright, 2019a).

Attention was first drawn to the glacial geology of the Miller Brook valley after publication of a short paper by Wagner (1970) who suggested that landforms in the Miller Brook valley and elsewhere in northern Vermont were the product of alpine glaciers that existed after the recession of the Laurentide Ice Sheet. Wagner's paper stimulated considerable discussion, mostly centered on whether or not the ridges Wagner interpreted as moraines were produced by a waning tongue of the Laurentide ice sheet or by a local glacier (Ackerly, 1989; MacClintock, 1971; Stewart, 1971; Wagner, 1971; Waitt and Davis, 1988).

Interpreting landforms in the Miller Brook Valley has always been confounded by the lack of detail shown on available maps. To overcome this problem a portion of the valley was mapped with a total station by two UVM students, M. Loso and H. Schwartz during the fall of 1995. This map and preliminary descriptions and interpretations of soil pits constructed by P. Bierman's geomorphology class are presented in Loso and others (1998). The mapping area was continued downstream by the author in the fall of 1996 and Wright and others (1997) used these maps and additional soil pits to conclude that the ridges described by Wagner (1970) are eskers overlain by a carapace of ablation till where they occur along the valley sides. Detailed surficial mapping of the valley occurred during the summer of 2017 on a LiDAR base map as part of a larger project mapping the Bolton Mountain Quadrangle (Wright, 2018b).

Stops A–F occur on trails maintained by and land owned by the Lake Mansfield Trout Club. While the Club has generously allowed both field work and field trips on their property in the past, please request permission before accessing the site. While most members do come to fish, many also make use of the Club's extensive trail network, including those we will use on this field trip. **Stops G and H** both occur on public land and the final two stops (**I and J**) occur within the abandoned parts of a gravel pit owned by the Town of Stowe.

Hiking Log:
Approx. Distance (km)
STOP A: LAKE MANSFIELD DAM **(-72.810659, 44.472195)**

0.0 The Lake Mansfield dam was first built by the Trout Club in 1901 to provide a habitat for trout. The club owns and maintains two smaller dams downstream from here. The valley head visible to the west is what Wagner (1970) interpreted as the cirque which sourced the post-Laurentide alpine glacier that deposited the "moraines" he described farther down the valley (Fig. 28). While the cirque-like form of the upper Miller Brook valley is less well-developed than cirques on higher-elevation mountains elsewhere in New England (Davis, 1999), this valley head and others still possess a distinctive bowl-shaped form difficult to produce by stream erosion alone. At the onset of each glacial regime within the Pleistocene alpine glaciers likely developed on the eastern slopes of Vermont's mountains, but existed for shorter periods of time and hence accomplished less erosion than those in the higher-elevation mountains before being overwhelmed by the ice sheet.

A lovely alluvial fan protrudes into the north side of the lake.

From the dam follow the Lake Mansfield Trout Club (LMTC) "White" trail east along the south side of the Miller Brook Valley. Follow this trail to the top of a meadow that looks east, down the Miller Brook valley.

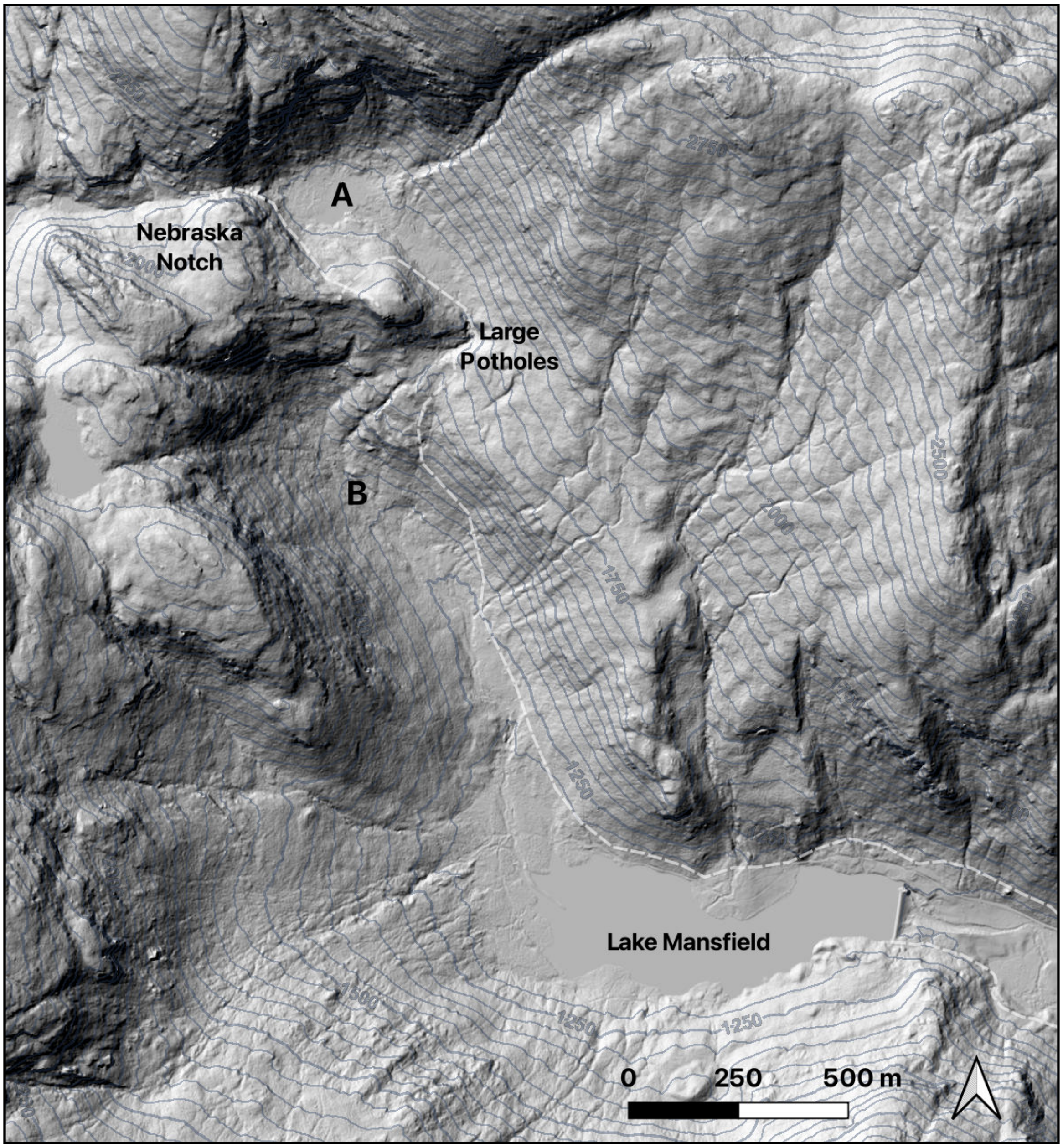


Figure 28: The upper Miller Brook valley from Nebraska Notch to Lake Mansfield. “A” and “B” mark two head-of-valley cirque-like landforms. Large, 7-10 m diameter potholes are located where the stream gradient increases abruptly. Brittle bedrock structures underlie pronounced lineaments. Contours in feet.

STOP B: ICE-CONTACT DEPOSITS DOWN-VALLEY FROM THE DAM (-72.809035, 44.471783)

0.2 The sediments underlying this meadow were exposed in test pits dug when the this area was being considered for the Club’s leach field (Fig. 29). The broad sloping ridge lying adjacent to Miller Brook consists of sand and gravel which contains one small kettle at its upper (western) end. In contrast, areas farther down the slope are underlain by silt and sand that is frequently faulted and disrupted by soft-sediment deformation. The ridge is interpreted to be the eroded remnants of ice-contact sediments deposited as a largely subaerial fan when the ice margin lay approximately where the Lake Mansfield dam currently exists. The silt and sand are ice-proximal lacustrine sediments deposited in the shallow water near the shoreline of Glacial Lake Winooski in this area, ~342 m (1,122 ft).

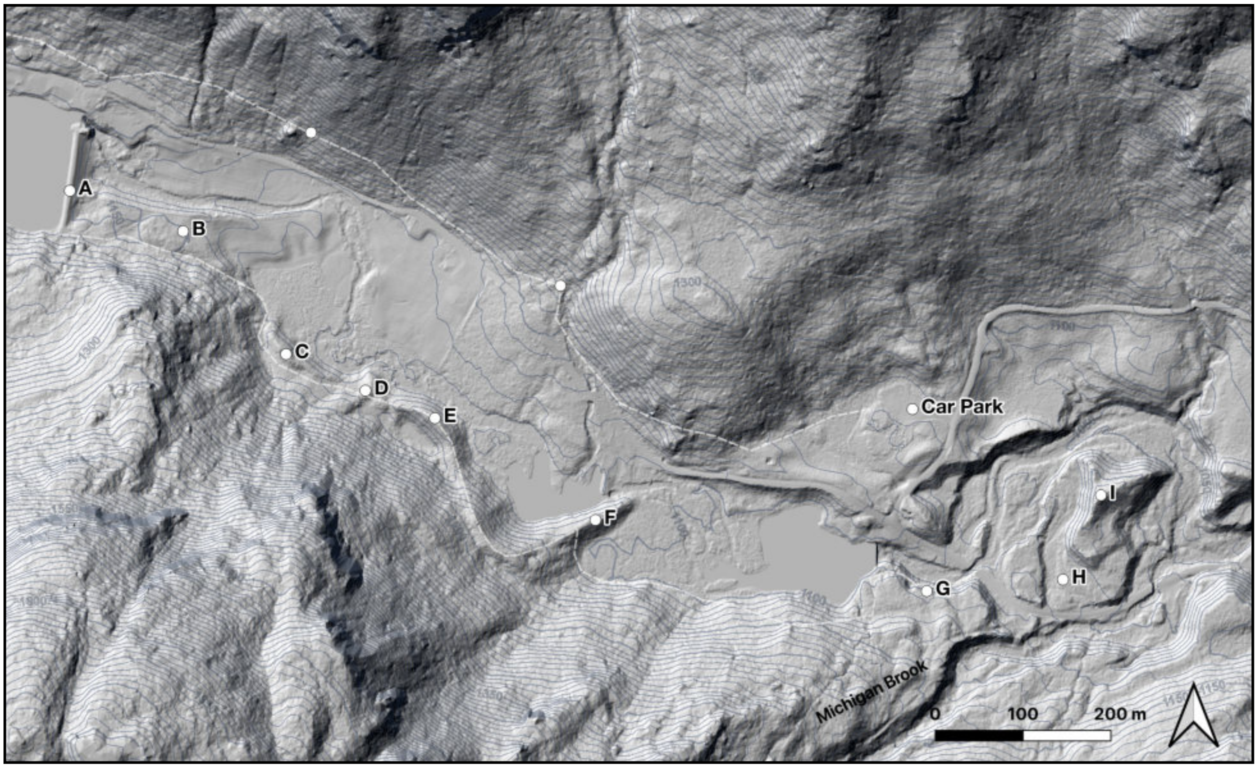


Figure 29: Miller Brook valley field stops along a ridge on the southwest side of the valley. A large alluvial fan lies on the northeast side of the valley, opposite sites D and F.

The broader paleogeographic picture at this time is one where the Laurentide ice sheet on the east side of the mountains had thinned and completely retreated well to the north of the Miller Brook valley. However, a tongue of ice from the much thicker ice sheet in the Champlain valley west of the mountains flowed through Nebraska Notch and down the Miller Brook valley. The sediments and landforms at this stop record the recessional position of this tongue of ice at or near the shoreline of Glacial Lake Winooski. As we walk down-valley we will be looking at sediments and landforms deposited perhaps only a few years earlier.

Return to the White trail and follow this down valley. The trail ascends a small intermittent stream valley lying between the north-facing valley side and a small ridge.

STOP C: RIDGE ALONG SOUTH SIDE OF VALLEY (-72.807555, 44.470527)

0.5 This and several of the following stops are along segments of the ridge that Wagner (1970) interpreted to be a moraine (Fig. 29). Here a distinctive ridge lies parallel to the valley side and stands with almost 10 m of relief from the valley floor. The LiDAR imagery clearly shows the geographic form and extent of the ridge. Two small alluvial fans, deposited from ephemeral streams draining the south side of the valley, have partially filled the gap between the ridge and the valley side.

Large boulders adorn the ridge and soil pits here and elsewhere along the ridge encountered a loose, unsorted mixture of angular rocks (Wright et al., 1997). The material littering the surface of these ridges and that accessible with hand tools is diamict. This diamict has been interpreted as till and a ridge of till parallel to a valley is logically interpreted to be a moraine (Wagner, 1970; Waitt and Davis, 1988). However, no equivalent ridge exists on the north side of the valley. If the ridge here is a lateral moraine, then (1) a correlative moraine was never deposited on the north side of the valley (unlikely!) or (2) the moraine that was there has completely eroded away.

STOP D: SHORT RIDGE SEGMENT OBLIQUE TO VALLEY SIDE (-72.806416, 44.470157)

0.6 This short ridge is also mantled with large rocks but is aligned southwest-northeast, i.e it angles down valley from valley side (Fig. 29). If this ridge is largely composed of till and is a moraine, it is an erosional remnant of a small recessional moraine deposited perhaps during a winter standstill. If it used to extend across the valley then the northern end of it may be buried by the large alluvial fan existing there and the middle segment has been eroded by Miller Brook. Another consequence of a moraine here is that it may have ponded meltwater from the retreating ice sheet and be the body of water the lacustrine sediments at Stop B were deposited in. An alternative interpretation is that the bulk of this ridge is actually composed of ice-contact sediments that were subsequently covered by diamict deposited during a debris flow down the steep valley sides although this scenario seems less plausible.

STOP E: RIDGE AND TRAPPED WETLAND; SOUTH SIDE OF VALLEY (-72.805420, 44.469877)

0.7 Stop E is on a very prominent ridge that stands more than 10 m above Miller Brook and arcs to the south (Fig. 29). During most parts of the year an ephemeral peat-filled pond occupies the space between the ridge and the side of the valley. Organics, sampled from a depth of 235 cm, were dated at 9280 +/-235 ^{14}C years (10,561 +/- 327 calibrated years BP) by Sperling and others (1989) indicating that the ridge is a primary geographic feature (not produced by erosion) standing along the valley side during all of the Holocene.

The elevation profile of this ridge varies considerably. Specifically, the elevation of the ridge between Stops C and E dips down where it crosses the moraine-like landform (Stop D) and then climbs above 350 m. Along the next stretch of trail the ridge first drops and then climbs considerably higher, reaching an elevation of ~364 m, before making a 90° turn to the northeast and steadily dropping to an elevation of ~345 m (Fig. XX). The varying elevation profile of this ridge is quite different from lateral moraines which tend to drop steadily in elevation down-valley mimicking the profile of the glacier that produced them.

STOP F: RIDGE BELOW MIDDLE TROUT POND; CENTER OF VALLEY (-72.803116, 44.468839)

1.0 After the ridge turns northeast it protrudes out into the center of the valley (Fig. 29). Wagner (1970 interpreted this segment of the ridge as an end moraine. Boulders are rare on this section of the ridge and even a shallow shovel hole reveals that the ridge is composed of coarse sand and rounded gravel. Similar test pits farther down valley all confirm that this ridge is composed of stream-transported gravel and not till. Consequently, this ridge is an esker and not a moraine. Inspection of the LiDAR imagery (Fig. XX) shows that the ridge is breached by Miller Brook, but continues down-valley and underlies the access road to the Trout Club.

The till that blankets the ridge where it lies along the side of the valley is most likely an ablation till that accumulated on top of the esker when the ice sheet melted. Much of this ablation till probably consists of till that slid via debris flows from the valley side as the glacier thinned. This is the till that has led many geologists to interpret these ridges as moraines. In the center of the valley the till cover on the esker is thin and patchy and is most likely derived from englacial debris.

The White trail drops off the esker and follows the south side of the valley skirting a wetland area immediately above the lowest dammed pond on the Trout Club property. After reaching the dam, follow an old road east along the south side of Miller Brook.

STOP G: GLACIAL LAKE WINOOSKI DELTA

This terrace consists of sand and gravel deposited as a delta in Glacial Lake Winooski by both Miller Brook and Michigan Brook (Fig. 30). In addition to the cut for the roadway, a recent landslide provides a good exposure of these sediments. The terrace here and underlying the car park lie at an elevation of ~342 m (1,122 ft) the local elevation of Glacial Lake Winooski,.

Follow this old road until it intersects the road just below the bridge over Miller Brook. Walk down the road (southeast) ~90 m and then enter the woods on the north side of the road. A short walk leads to an abandoned stream channel and a series of stepped terraces above the channel.

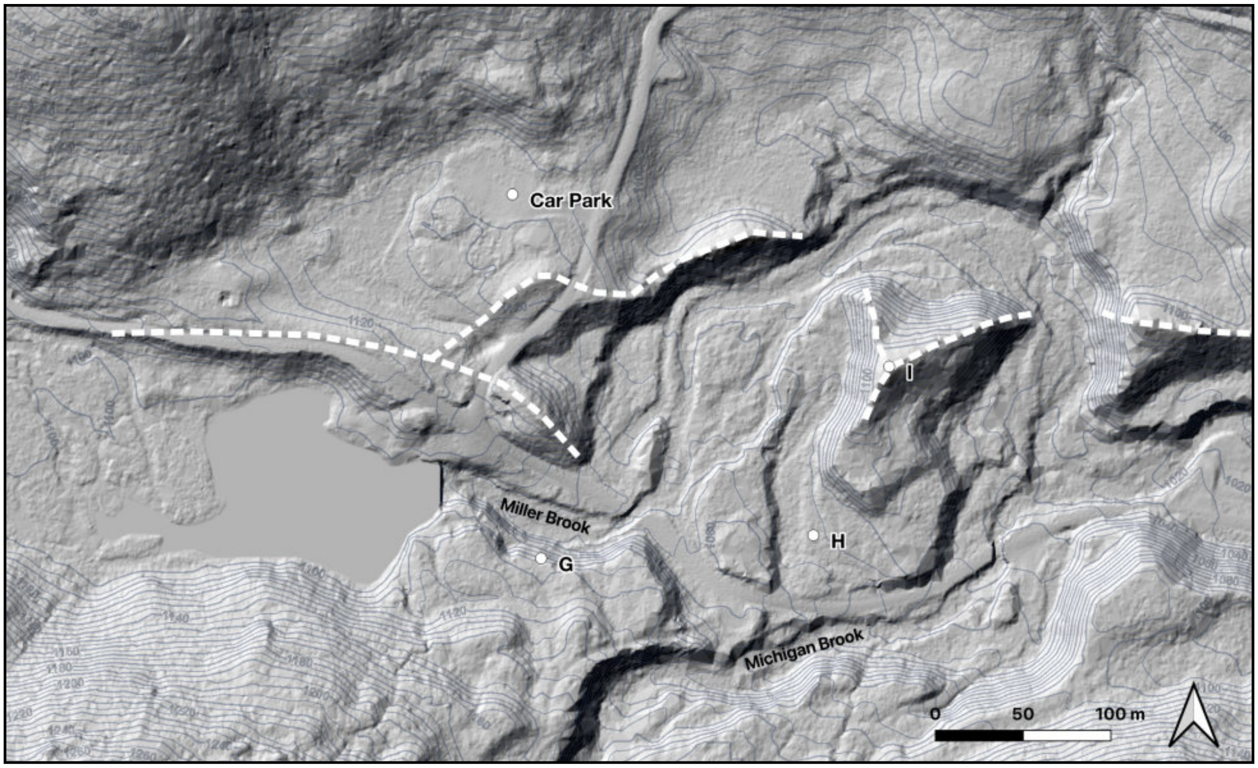


Figure 30: Detailed map of esker ridges and abandoned stream channels at the confluence of Miller and Michigan Brooks. The terrace at Stop G and the Car Park is a Glacial Lake Winooski delta deposited by both brooks. The esker (highlighted with whited dashed line) bifurcates in this region and was first largely buried by both subaqueous fan and delta sediments and later re-exposed by stream incision following the drainage of Glacial Lake Winooski from the valley.

STOP H: ABANDONED CHANNELS; INCISED GLACIAL LAKE WINOOSKI DELTA COMPLEX
(-72.796418, 44.468246)

1.7 This stop is located near the confluence of Michigan Brook and Miller Brook (Fig. 30). An abandoned Michigan Brook stream channel cuts N-S across this area. The channel is bordered by several adjacent stream terraces giving this landscape a stair-step topography. Immediately to the west (south of the car park) the esker ridge bifurcates and has been eroded by both Miller and Michigan Brooks. Stream terraces and channels situated well above the valley floor indicate that the eskers were largely buried by sediments (i.e. base level was high) prior to stream incision. Some of these sediments were deposited as subaqueous fans as the ice sheet retreated across this area. After the ice sheet retreated up the valley both Miller Brook and Michigan Brook formed deltas which merged together in this area and largely buried all but the highest parts of the esker system (Fig. 30). The complex network of stream channels were incised after glacial Lake Winooski drained ~13,800 years ago.

Climb the steep slope to the north. After crossing one of the high terraces, continue climbing up the esker ridge to its high point.

STOP I: INTERSECTING ESKERS **(-72.795880, 44.469112)**

1.8 This is the intersection of three esker ridges in the middle of the valley with Miller Brook flowing around this point to the west, north, and east and Michigan Brook passing to the south (Fig. 30). There is a 17 m drop in elevation to the abandoned channel to the north. The high esker ridge arcs from N-S to east where it has been cut by Miller Brook and then continues down the valley. The southern extension of this esker has been largely eroded away, but likely used to turn back to the west where it connected to northeast of the dam (Fig.

30). The esker extending to the north is also cut by Miller Brook and continues as a sinuous ridge below the parking area. This esker is lower in elevation and appears to be cross-cut by the higher esker. While this esker may be older, it's also possible both subglacial tunnels were active at the same time and this northern loop was just a smaller tunnel.

2.3 Return to the Lake Mansfield Trailhead car park.

<u>Stop B4 (J, K): Nebraska Valley Solar Farm and Gravel Pit</u>

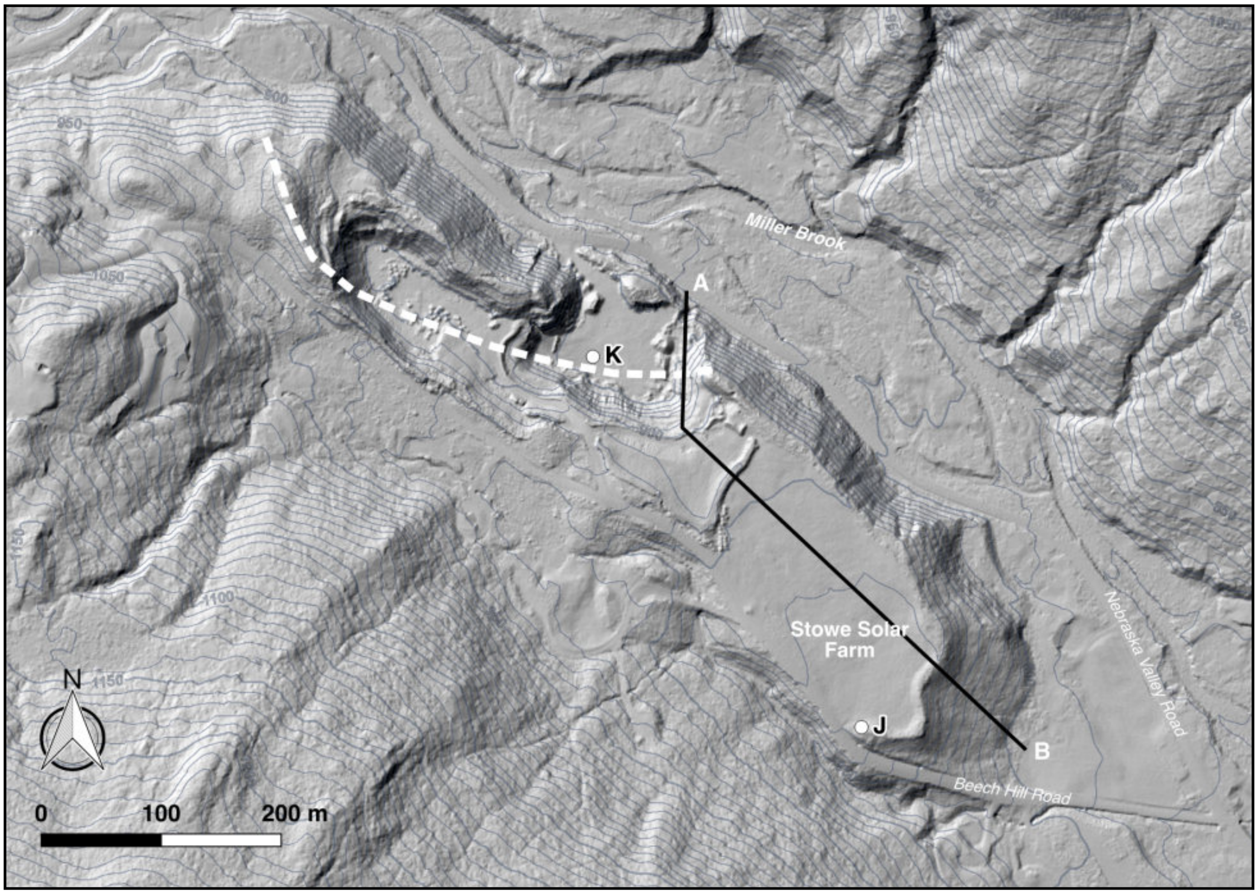


Figure 31: Map of the Miller Brook gravel pit and adjacent solar farm. White dashed line is the crest of the Miller Brook esker. The Stowe Solar Farm lies in an old gravel pit consisting of old Miller Brook alluvium. These sediments unconformably overlie Glacial Lake Winooski varved silt/clay and are graded to the elevation of Glacial Lake Mansfield 1.

STOP J: NEBRASKA VALLEY SOLAR FARM (-72.769546, 44.461475)

The Stowe Solar Farm is built on a terrace that was formerly a meadow and then a gravel pit (Fig. 31). The berm around the perimeter of the pit shows the former elevation of the terrace (slopes gently SE 275–270 m, ~900–885 ft) and indicates that only several meters of gravel have been removed. The gravel here unconformably overlies varved silt/clay deposited in Glacial Lake Winooski. The gravel that formerly covered this terrace is most likely old Miller Brook alluvium deposited on the floor of Glacial Lake Winooski after the lake drained (Figs. 31, 32). Miller Brook meandered across this terrace with a gentle gradient without incising its channel further into the underlying lacustrine sediments. This implies that the brook was close to base level which was the local elevation of Glacial Lake Mansfield 1, the lake that formed after Glacial Lake Winooski partially drained (Fig. 9).

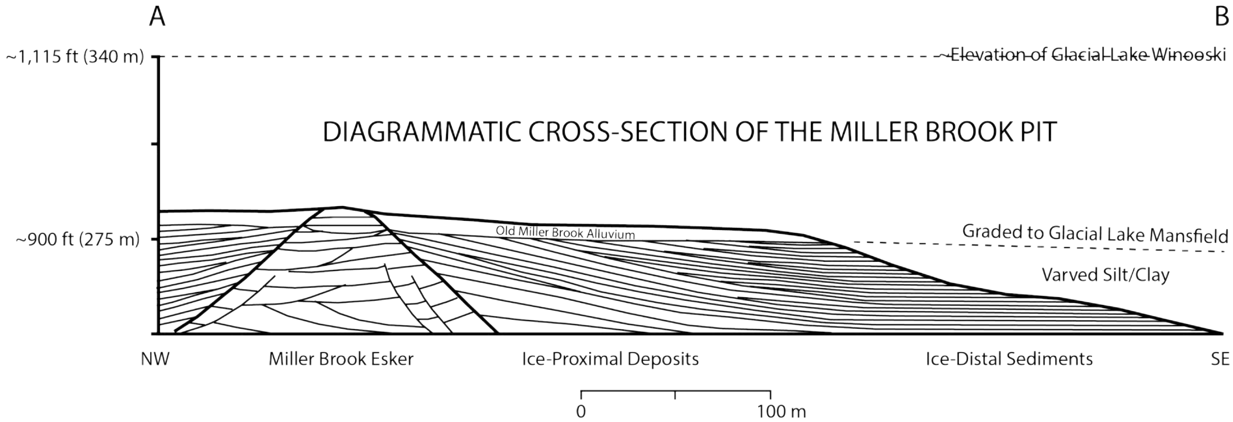


Figure: 32: Diagrammatic cross-section across both the gravel pit and solar farm (see Fig. XX for location of section). Esker sediments are overlain by a fining up sequence of sediments deposited in Glacial Lake Winooski as the ice sheet retreated up the valley. After Glacial Lake Winooski partially drained, Miller Brook began depositing alluvium on the floor of the former lake before flowing into Glacial Lake Mansfield.

STOP K: MILLER BROOK GRAVEL PIT

The Miller Brook esker extends discontinuously from Lake Mansfield down the valley as far as this gravel pit (Fig. 31). The objective of this stop is to observe sediments marking an almost complete transition from an esker tunnel environment to the relatively quiet waters of Glacial Lake Winooski.

The undisturbed esker ridge is prominently exposed in the woods to the northwest and can be traced across the pit floor (see dashed line in Fig. 31), although it's being actively removed and distributed across various parts of Stowe. Recent exposures of the esker sediments show bedding dipping moderately to gently to the southeast and consisting largely of coarse boulder/cobble gravel interlayered with beds of coarse sand and pebble gravel. Small scale faults, most with a normal sense of slip, are common along the exposed border of the esker.

The esker is overlain by a fining-upward sequence sediments ranging from pebble gravel close to the esker to very fine sand, silt, and clay away from the esker (Fig. 32). These sediments are best exposed at the northwestern end of the pit. Dropstones are common as are distinct, thin silt layers within the sand horizons. Varved silt and clay[3] layers lie at the top of the section, farthest from the esker. Spectacular folds and faults occur within these varves and were likely produced when these soft yet coherent sediments slid off the northeast side of the esker into deeper water. The projected elevation of Glacial Lake Winooski over this part of the Miller Brook valley is ~340 m (1,115 ft) (Fig. 16). Therefore, as the ice sheet retreated, the crest of the esker (elevation 300 m) was beneath ~40 m of water and and areas on either side in much deeper water. Coarse sediments ejected from the esker tunnel were deposited as subaqueous fan sediments close to the tunnel mouth. As the ice sheet retreated farther up the valley, the average size of sediment originating from the esker tunnel gradually diminished producing the fining-upwards sequence of sediments currently draping across the esker (Fig. 32).

[3] Senior thesis work by Derr (2011) has shown that the winter "clay" layers in varves collected at the Waterbury Reservoir (~8.5 km south of the Miller Brook pit) actually consist of fine silt size particles consisting largely of the minerals muscovite, clinochlore, and chlorite, common mica minerals in the surrounding rocks.

The next three stops all occur near the confluence of the Mad River with the Winooski River in Middlesex (Fig. 33). Areas adjacent to the rivers are underlain by an complex assortment of sediments deposited in both ice-contact and lacustrine environments.

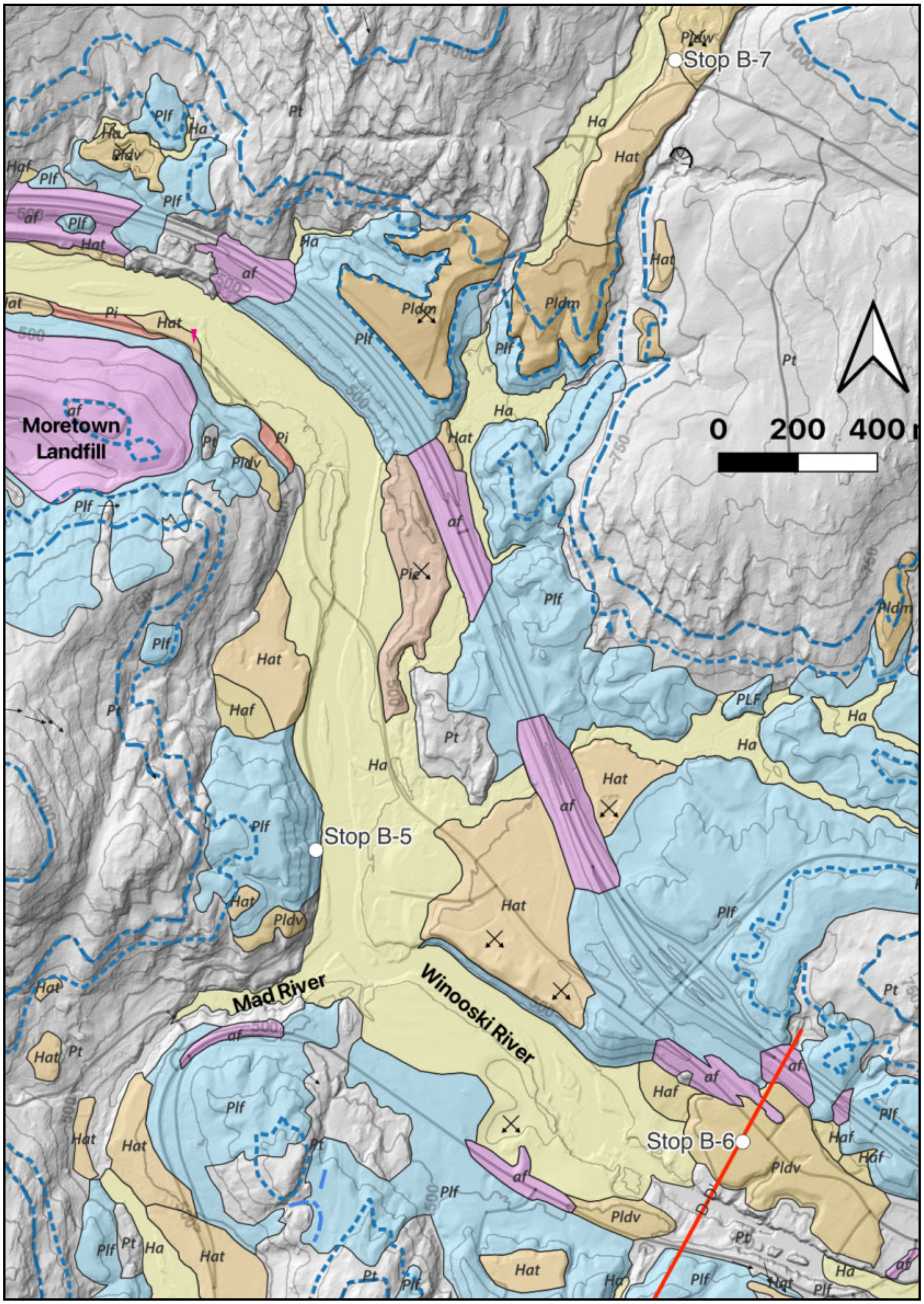


Figure 33: Portion of the Middlesex Surficial Geologic Map (Wright et al., 2023) showing the geology around the confluence of the Mad and Winooski Rivers. Dashed blue line show the projected elevations of Glacial Lake Winooski whereas dash-dot lines outline Glacial Lake Mansfield 1 & 2. Pt: Till, Pie: Esker, Plf: Lacustrine fine-grained sediments, Pld: Lacustrine delta, Hat: Alluvial Terrace, Haf: Alluvial Fan, Ha: Alluvium, af: Fill. Cross-section D-D' shown with red line.

Stop B-5: Lovers Lane Landslide: Subaqueous Fan Sediments

Many landslides formed or were reactivated during the July 2023 floods in northern Vermont. A relatively small landslide along Lovers Lane exposes a small section of lacustrine sediments that wasn't visible when this area was recently mapped (Fig. 33; Wright et al., 2023b). Layers of medium to fine sand and gravel dip moderately north and contain a variety of small scale sedimentary structures. Sediments farther up the steep slope consist mostly of fine-grained glacial lake sediment, i.e. silt/clay/very fine sand. It therefore seems likely that these sediments were deposited in a subaqueous fan as the ice sheet retreated WNW down the valley. Gravel deposited on bedrock in a disused pit at the State Garage on the opposite side of the valley may be part of an esker.

Evidence presented at the next stop indicates that a Glacial Lake Mansfield delta extended down the Winooski River valley at least as far as Middlesex Village.

Stop B-6: Camp Meade Terrace, Middlesex: Buried Winooski River Channel/Tromino Survey

Middlesex village is built on a terrace that is the former location of a 1930's era CCC camp called "Camp Meade." The modern channel of the Winooski River lies directly south of the village where the river is floored by bedrock and dammed at a substantial water fall. The below cross-section is controlled by surficial mapping, bedrock exposures, a single water well, and a Tromino passive seismic survey (Fig. 34; Wright et al., 2023b). The water well ends in coarse-grained sediments some distance above the bedrock. This well and the Tromino data define a deeply buried bedrock channel directly beneath this terrace and some distance north of the modern channel. Sediments recorded in the water well log fine up and likely record lacustrine deposition as the ice sheet retreated down the river valley. Coarse sediments directly underlie the terrace which lies at the projected elevation of the Coveville stage of Glacial Lake Vermont. Following the drop in lake level first from Glacial Lake Winooski to Glacial Lake Mansfield 1&2 and then from Glacial Lake Mansfield 2 to Glacial Lake Vermont, the Winooski River may have deposited these terrace gravels as alluvium on the newly exposed lake floor or as a thin delta that rapidly extended across the shallow lake.

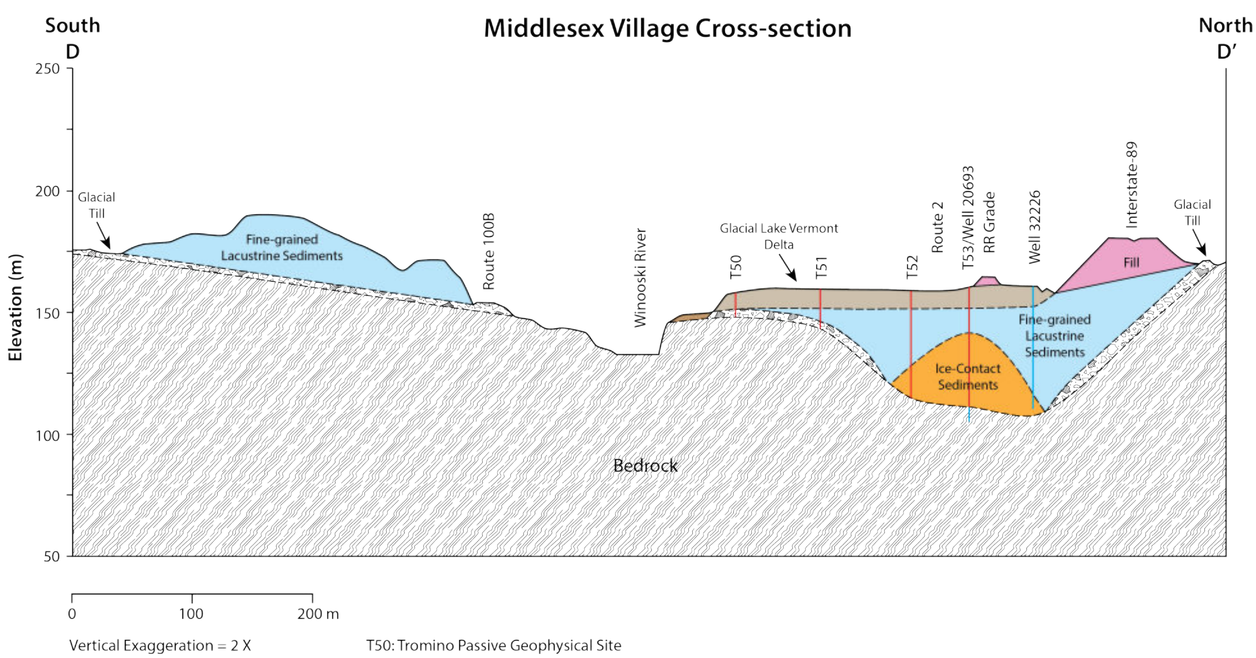


Figure 34: Cross-section across the Winooski River valley at Middlesex Village (see map for location). An old buried channel of the Winooski River is clearly outlined by both the water well and Tromino passive seismic data. Water well logs within the buried channel record extensive sand and gravel deposits lying beneath lacustrine sediments. These coarse sediments are interpreted as ice-contact sediments likely deposited in an esker. These in turn are overlain by a coarsening up sequence of sand and gravel interpreted as a Glacial Lake Mansfield delta.

Stop B-7: Middlesex Town Pit

This pit is located on the north side of the Winooski River, south-southeast and immediately below the Middlesex Notch spillway of Glacial Lake Thatcher, a small short-lived lake dammed on the west side of the Worcester Mountains (Dunn et al., 2011; Springston and Dunn, 2006). What can be seen in this pit changes fairly significantly as sand and gravel is removed. An important problem here concerns the extent of ice in the Winooski and Thatcher valleys, or conversely the extent of glacial Lake Winooski, when the sediments here were deposited. We have envisioned a lobe of ice in the Winooski valley, with sediments derived locally and from Middlesex Notch, and deposition taking place onto or against the flank of this ice, with some degree of local meltwater influence; however, we welcome thoughts on sedimentary environments of deposition represented here.

The entrance to the pit lies at about 840 ft asl, and deposits here rise to ~1000 ft asl, which is the projected glacial Lake Winooski shoreline, and in fact they rise above this by several 10's of feet, grading down to the 1000 ft elevation. Between Site 3 and the Middlesex Notch spillway there are sandy gravels preserved on lodgement till. On the topographic map a prominent surface extends SSW, below the pit location Auger holes in this surface revealed bedded sand, but access to the area is restricted and further examination is not possible. The surface extends to ~700 ft asl, suggesting it is graded onto deposits of glacial Lake Winooski, or Mansfield I.

If this is so, then was there ice in the Winooski valley when most of the sediment in the pit was deposited, and then the relatively short-lived glacial Lake Winooski formed, and finally the end of deposition occurred as sediment was graded down to the glacial Lake Winooski bottom? Or, have we given too much credit to a sediment supply from Middlesex Notch, and in fact some or most of the sediments here derive from ice-contact deposition as ice retreated down the Winooski valley and glacial Lake Winooski expanded into this area? Or, what other possibilities are there?

There is an upper pit to the north, above the level we are visiting. In this area we have found several unusual but vertically small sections. One example includes 2 ft of deformed (collapsed), poorly sorted pebbly sand overlain by 3 ft of moderately sorted, pebbly medium-coarse sand that contains up to 1 foot diameter hard, silty diamict clasts (interpreted as till rip-up clasts), which is in turn overlain by 4 ft of fine-medium sand stringers and lenses around 0.5-1.5 foot diameter silt-sand-pebble high density/firm diamict. The diamict clasts are tabular and appear as a fractured unit, floating in a sandy matrix, almost resembling a loose stone or brick wall in their packing. This deposit lies at or above 1090 ft asl, or well above glacial Lake Winooski. What is the origin of this unit? Is it another till, representing a readvance, or a debris flow? Other ideas?

Sunday June 9, 2024: Marshfield/East Barre Area

The goals of Sunday's stops (Fig. 35) are to examine and discuss typical glaciolacustrine deposits of glacial Lake Winooski, and to examine and discuss deposits related to what we interpret as a local readvance of the ice margin, probably early in the period of glacial Lake Winooski. We will begin by examining the present geomorphology of lake bottom deposits in the main lake basin and then, after a brief look at a modern alluvial fan, we will work our way into small sub-basins in which proglacial lakes were trapped by the readvancing ice margin.

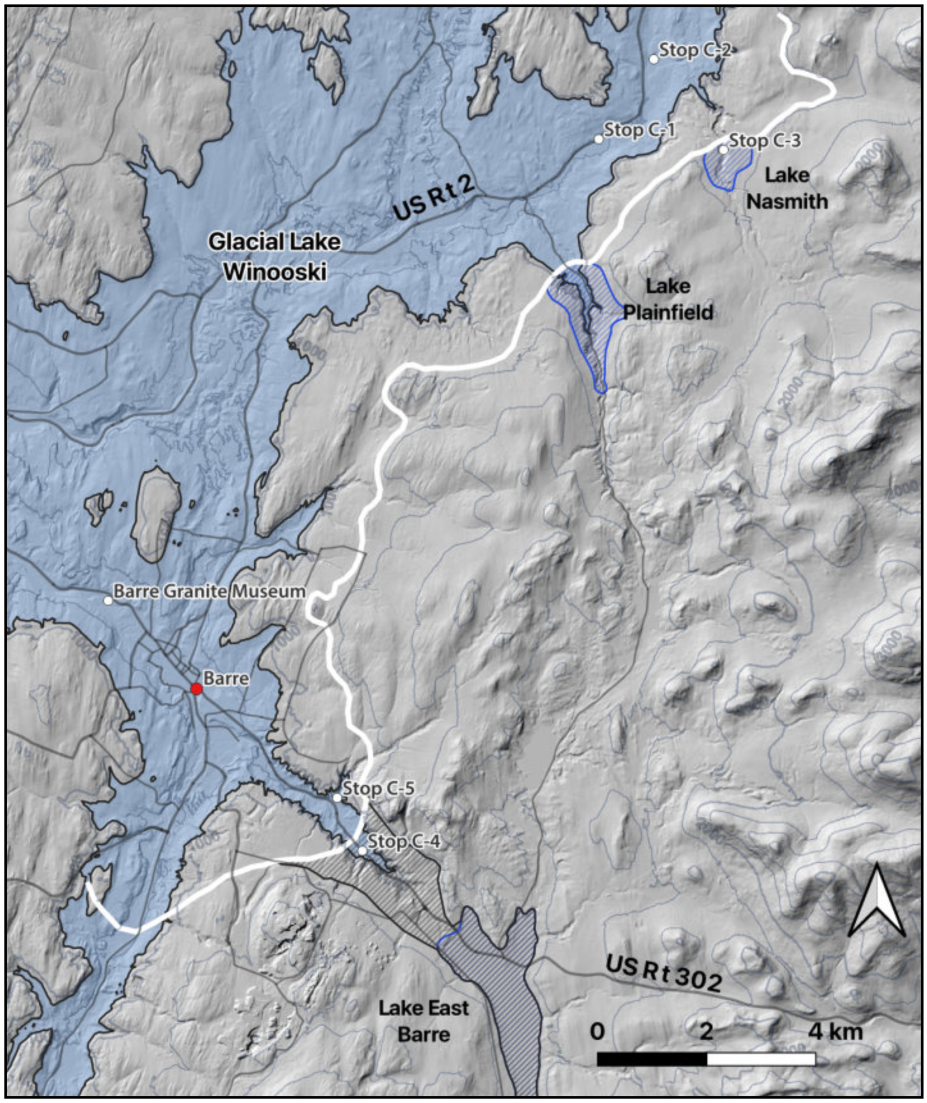


Figure 35: Map of the area of Sunday's field stops. Upper glacial Lake Winooski is shown in pale blue and an inferred readvance ice margin that would produce small isolated proglacial lakes is shown with a white line, although the readvance must have extended more southeast to produce all of the deposits seen at the field stops. The three ice-marginal lakes created when ice blocked sub-basin outflow are shown with hatch pattern and include glacial lakes Nasmith, Plainfield, and East Barre.

Across valley bottoms of the Winooski basin, distal glaciolacustrine deposits are dominated by plane bedded fine-medium sand, typically displaying ripples, and rhythmites of sand-silt-clay, often as varves. Sediments were carried into the ice-dammed lake basin by overflows, interflows, and underflows and deposited by traction, saltation, and suspension settling. Distal lacustrine sediments are 30+ meters thick in the main valley bottoms, as recorded in borehole data and limited outcrops. Common proximal glaciolacustrine deposits of glacial Lake Winooski include coarse-grained subaqueous outwash fans and deltas. Excellent examples can be found around the lake perimeter and are often exposed in sand-gravel pits. They can be recognized by their coarse modal grain size, fining in the basinward direction, cross-stratification and graded beds, and basinward-dipping beds.

While lake subbasins typically contain deposits similar to those of the main Winooski basin, several subbasin sites have been found the middle and upper Winooski basin where the stratigraphic sequence records what we interpret as a readvance of the ice margin (Fig. 36). Evidence at these sites generally consists of deformed lake beds and two-till sections (Larsen, 1999a; Larsen, 1999b; Larsen, 2001; Larsen et al., 2003b; Wright, 2015a). Larsen (2001)used this stratigraphy and age to propose the Middlesex Readvance which appears to be coeval with the readvance recorded at the Comerford Dam site in the Connecticut River basin (Thompson et al., 2017). More recently we have described readvance sites in Nasmith Brook, Great Brook, and Jail Branch valleys in upper glacial Lake Winooski and it is these that we report on here.

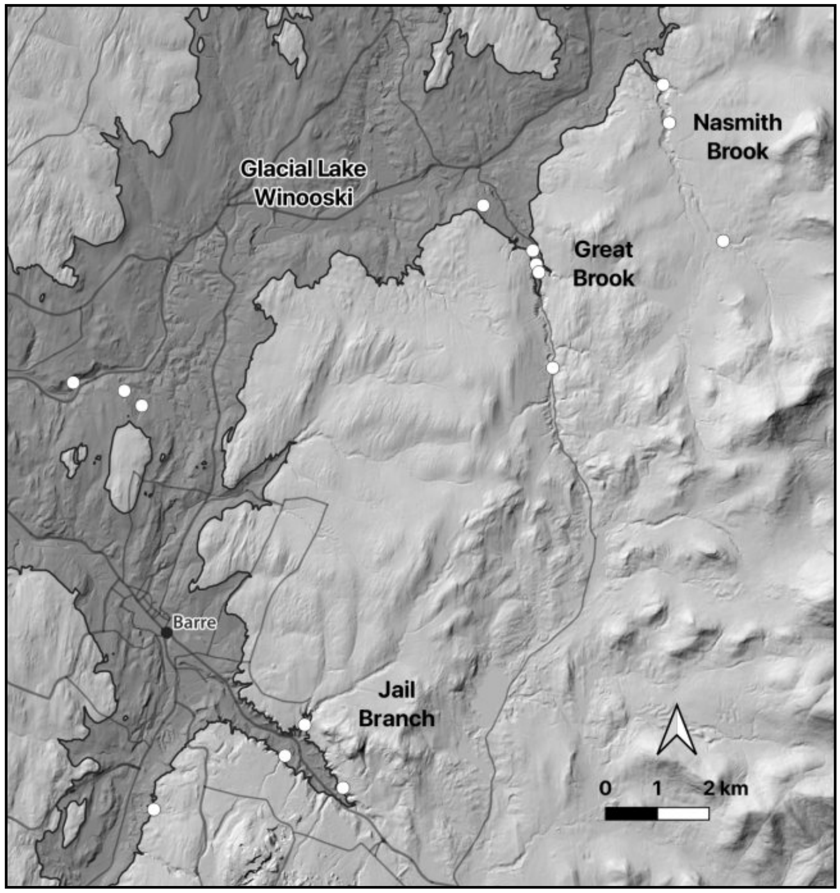


Figure 36: White dots show locations where diamict layers within sections of glaciolacustrine sediment may result from ice readvance.

The sections in the valleys of the Nasmith, Great, and Jail streams generally comprise a basal diamicton that lies upon bedrock and has a clast fabric indicating ice flow to the SSE that is overlain by variable thickness of glacio-lacustrine sediments that is in turn overlain by a thick, compact and massive diamicton. Larsen (unpub) first measured and described one of these sections in the Jail Branch south of Barre and interpreted it as a product of a proglacial lake, his preglacial Lake Merwin, that formed in front of the advancing late Wisconsinan ice and that was in turn overtopped by the ice, with a lodgement till capping the sequence (Larsen, 1999a). The basal till found elsewhere was not observed in this particular Jail Branch section but the lower approximately 1/3 of the section is covered in colluvium. We re-interpret the Jail Branch section, and add several others, to be a product of initial Wisconsinan ice advance and deposition of a basal till followed by ice retreat and proglacial lake formation and culminating in a short-term ice readvance that formed local, short-lived proglacial lakes in the basins that drained back against the ice margin (to the northwest). In addition to a readvance till some of these sites contain a thick deposit of stratified diamicton that is suggestive of deposition at the grounding line, and the glacio-lacustrine sections contain extensive soft-sediment deformation, and both cohesive and noncohesive remobilized deposits that we interpret as a function of glacio-tectonism of lake bottom deposits. Our stops will serve to demonstrate that the deposition in these subbasins was dynamic and highly variable, and strongly influenced by the advancing ice margin, both as a sediment source and as a deformational mechanism.

The stops on Sunday are found mainly on the Marshfield, Plainfield, Barre East, and Knox Mountain 7.5-minute quadrangles. Underlying bedrock is Devonian in age and includes the Waits River Fm., Gile Mountain Fm., Knox Mountain Pluton and the Barre Pluton (Ratcliffe et al., 2011). This portion of central Vermont has been the scene of several surficial geologic mapping projects in recent years. The surficial geology of the Plainfield quadrangle was mapped by Springston (2011), the Barre East 7.5 quadrangle was mapped by Springston (2018), and most of the Marshfield quadrangle and the northwestern part of the Knox Mountain quadrangle were mapped by Springston and Kim (2008). To the west of the study area, the Barre West quadrangle was mapped by Wright (1999b) and the Montpelier quadrangle was mapped by Larsen (1999b).

Stop C-1: Martin Bridge Park, Route 2, Winooski River valley, Marshfield (44.4885°N, 72.4070°)

This publicly owned site provides an overview of the Late Pleistocene and Holocene history of the Upper Winooski valley (Fig. 35). The surficial deposits in the valley consist of a sequence of Holocene stream and alluvial fan deposits overlying Pleistocene deposits of glacial Lake Winooski, which, in turn, overly till or, in a few locations, ice-contact sand and gravel (Springston, 2011). At this site till is exposed on the steep hillsides to the west and east of the broad valley bottom. The projected glacial Lake Winooski shoreline is at about 298 m (978 ft) asl. The modern floodplain is at 228 m (748 feet) and adjacent terraces are underlain by distal glaciolacustrine deposits consisting of laminated silt and silty clay overlain by silty fine sand and fine sand. At this site we will visit a terrace at about 245 m (800 feet) that appears to be a remnant of the lake bottom. At the southeast side of the terrace the lake deposits progressively thin above a bouldery till and pinch out at a prominent break in slope at the edge of the woods. The glacial lake shoreline itself is about 350 m farther to the southeast. Borings and water well records reveal that the surficial deposits in the valley bottom vary from less than 3 meters on the slopes on the west side of the valley to greater than 36 meters in the central and eastern parts of the valley bottom.

Site C-2: Alluvial fan at mouth of Nasmith Brook, Winooski River valley, Marshfield (44.2994°N, 72.3965°W)

This site is an active alluvial fan formed by Nasmith Brook as it flows steeply out of the hills to the southeast to join the Winooski River (Fig. 35; Springston, 2011). The fan was active during several high-flow events in the last two decades. Like Site C-1, this site is in the exhumed valley bottom of the Winooski River. Elevation at the confluence with the Winooski is 232 m (760 ft) asl. On the south side of Nasmith Brook the Twinfield Union School sits on a terrace at ~242 m (~794 ft) asl. The terrace is a Pleistocene alluvial fan formed immediately after drainage of glacial Lake Winooski before the lake bottom deposits were incised by the river. The shoreline of glacial Lake Winooski is at 300 m (984 ft) asl.

The purpose of this stop is to look at evidence of modern flood scour and deposition on the fan due to the large flood of July 2023. The athletic fields are built on the alluvial fan and although they have experienced flooding in the past, the 2023 episode appears to be the most severe in the last 25 years.

Stop C-3: Nasmith Brook valley. Glacial Lake Nasmith (44.2886°N, 72.3793°W)

Nasmith Brook flows north into the Winooski River (Fig. 35). The base of the section at this stop is approximately 35 m higher than the level of glacial Lake Winooski, which is at 299 m (981 ft) asl. As ice advanced south up the Nasmith Brook valley it blocked outflow, creating an ice-marginal lake that we have called glacial Lake Nasmith (Dunn et al., 2018; Dunn and Springston, 2019; Dunn et al., 2019). At this stop we can examine silt-sand-gravel deposits of the lake, which reached a minimum elevation of ~360 m (1180 ft) asl, and an overlying diamicton interpreted as a lodgement till. The Wilber property lies along the top of the large exposure on the northeast side of Nasmith Brook. This ~30 m high exposure was heavily eroded during a flash flood on the evening of May 29, 2011 (Fred Wilber, personal communication, 2019) and was subsequently modified by more recent flash floods. As we cross the property note the abundance of boulders—these represent the capping till of the sequence here. We will descend the exposure and visit our measured section at the northern end (Figs. 37, 38). In general, this section consists of proximal over distal glaciolacustrine deposits that are in turn unconformably overlain by post-glacial fluvial deposits. Across the rest of the exposure the glaciolacustrine deposits are overlain by a thick till cap.

Section Details

The lower part of the measured section contains distal glaciolacustrine deposits with local gravel-rich bodies of subaqueous outwash, perhaps formed by sediment gravity flows, that are the initial record of readvancing ice in this location. Higher in the section, beds of silty fine sand appear internally dismembered and dewatering structures become common. These appear to be semi-cohesive mass failure deposits. The upper portion of the section contains several examples of sediment gravity flows (look for faint cross bedding) and mass failure slide deposits, including two silt horizons that can be examined for a discussion of their origin. A listric fault with extensional features in the hanging wall is present in the upper portion as well, but its origin is not clear. Ripple drift cross lamination and cross beds in this section all display a south paleoflow, that is, up the valley. Overall, the stratigraphy records the advancing ice front which is the principal sediment source. In addition, the impact of the ice on the lakebed seems to have dislodged existing deposits, remobilizing materials as noncohesive and cohesive mass failures.

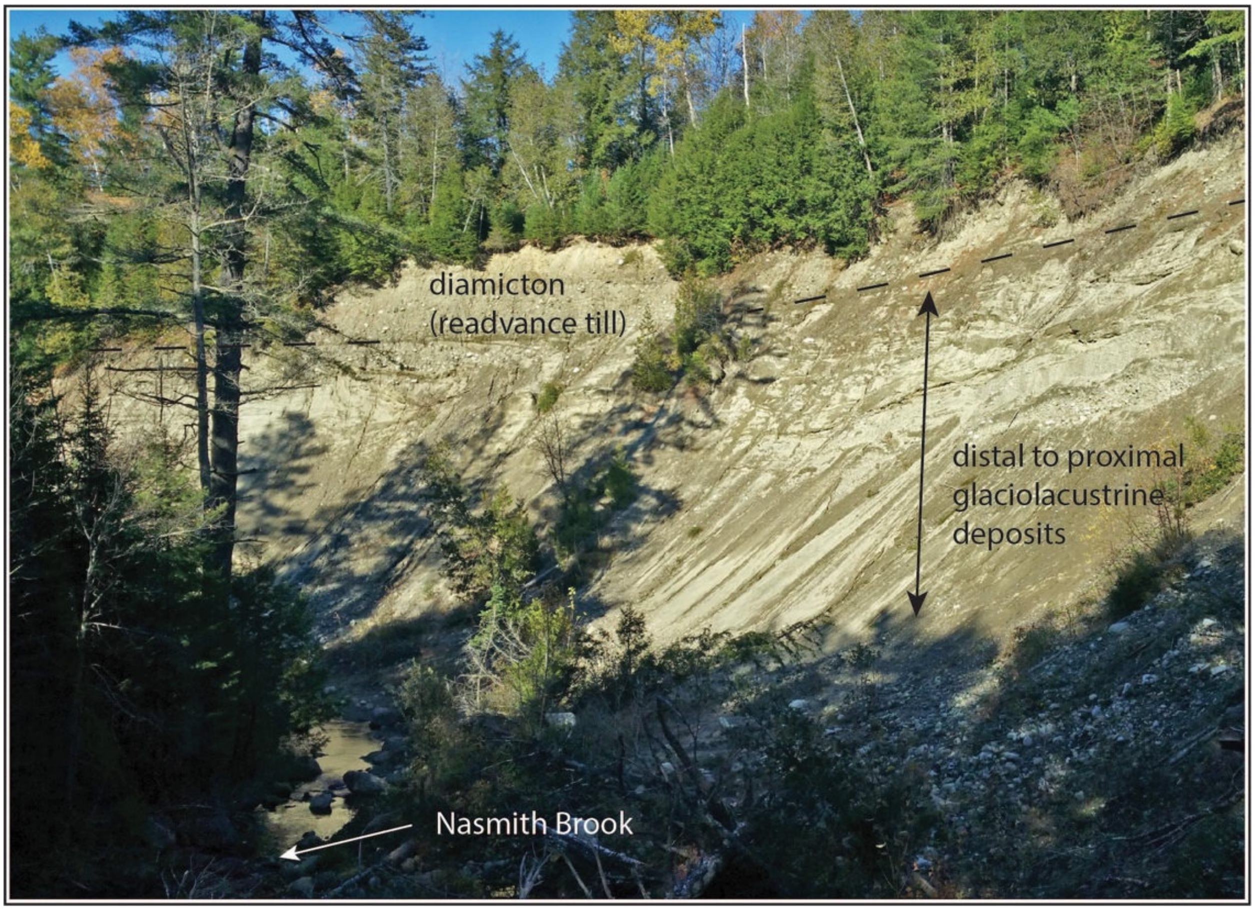


Figure 37: Section on Nasmith Brook, Wilber property, view approximately North. Glaciolacustrine sediments, from various depositional mechanisms, coarsen up section, the result of an advancing sediment source, but perhaps also of a shoaling lake, and the section is capped by thick diamicton interpreted as lodgement till deposited when readvancing ice overrode this position. Photo K. Rogers.

South of the measured section large channel-form gravels can be observed in the upper part of the outcrop. The gravels, and most of the exposure, are overlain by a massive, thick diamicton that we interpret to be a readvance till. Whether the gravels are relatively deep-water, subaqueous meltwater deposits or outwash deposits of a shoaling lake has not been established. We do not know the rate of rising water in the lake and thus the water depth during the deposition of gravel, and till, is not clear.

Travel Notes:

⇒ As we drive between Stops C-3 and C-4 we will travel up the Great Brook valley on Brook Road. The Great Brook watershed has experienced several severe floods in the past five decades and in consequence has experienced multiple episodes of channel downcutting and slope failure. The causes and mechanisms of fluvial adjustment and the surficial geology have been investigated in a series of studies (Barg and Springston, 2001; Springston, 2017; Springston and Barg, 2002; Springston et al., 2023; Springston and Thomas, 2014, 2018).

⇒ After crossing the watershed divide we will be in the Jail Branch watershed. Once we have turned north on Rt. 302, note the large dam on your left that was constructed in the 1930s to control flooding on the northwest flowing Jail Branch, which flows into Barre.

⇒ After the dam note a terrace surface on your right (east). After the Rt. 302 traffic circle there are several exposures in the terrace that can be seen while driving. The terrace is marked well by the growth of pines along its top. Stop C-4 will be an opportunity to discuss the terrace.

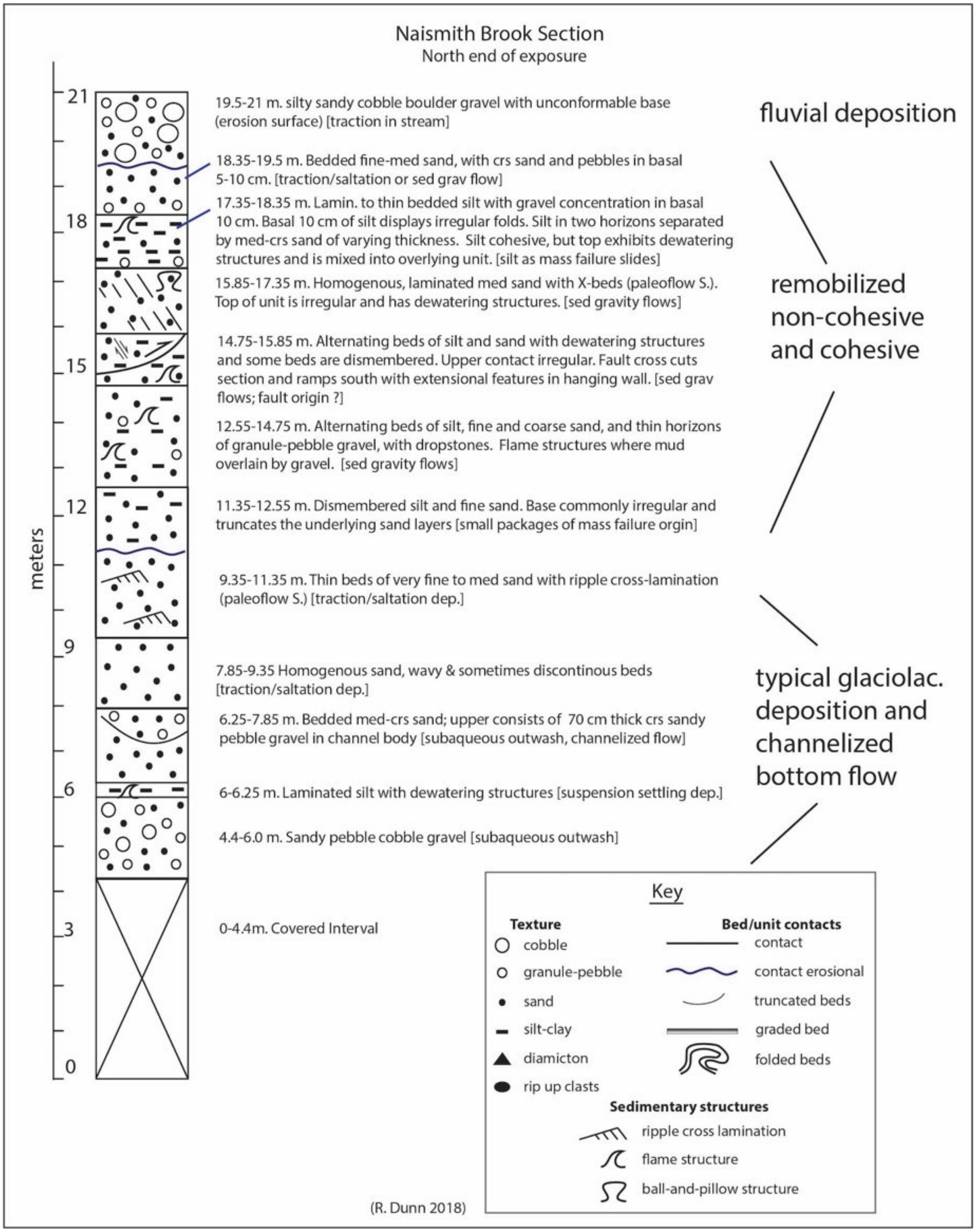


Figure 38: Measured section at north end of exposure at Stop C-3, Nasmith Brook. Glacial Lake Winooski was ~35 m lower than the base of the section. Measured by K. Rogers, R. Dunn, and G. Springston.

Stop C-4. Overview of Jail Branch valley from lot at Village Cannery warehouse and Swish Kenco building, Rte. 302. Glacial Lake East Barre (44.1703°N, 72.4627°W)

The Jail Branch flows northwest into Stevens Branch, which in turn flows into the Winooski River (Fig. 35). At this stop we can observe the large terrace on the northeast side of the valley, at approximately 330 m (~1080 ft) asl (Fig. 39), and a similar terrace occurs on the southwest side of the valley. The level of glacial Lake Winooski here was about 280 m (940 ft) asl, or ~50 m below the terrace surface. Down valley this difference diminishes, due to the gradual decline of the terrace down valley, and the difference at Stop 5 is ~16 m. The shape of the bedrock valley under Jail Branch is not fully established but well records indicate a thickness of unconsolidated material exceeding 30 m (100 ft) below the terraces, and generally 6 m (20 ft) in the valley bottom. This unconsolidated material is presumably all Pleistocene glacial fill and occurs as a flat-topped wedge that thickens down valley (to the northwest). Post-glacial stream erosion has created a prominent terrace and steep-walled drainage.

Ice flow during readvance blocked outflow from the basin and created an ice-marginal lake that we have called glacial Lake East Barre (Fig. 35; Dunn et al., 2018; Dunn and Springston, 2019; Dunn et al., 2019). In outcrops along the valley bottom the Pleistocene units comprise distal lacustrine deposits, especially silty rhythmites. Those sediments below glacial Lake Winooski level could have been partially deposited in Lake Winooski and partially in glacial Lake East Barre. Outcrops in the upper part of the terrace feature, like those we can see at this stop, are well above glacial Lake Winooski level and must have been deposited in glacial Lake East Barre. Materials are coarse-grained, proximal glaciolacustrine deposits overlain by a dense diamicton that is up to 10 m thick and that we interpret as lodgement till (Fig. 40). This distal to proximal sequence fills the entire valley from Barre to the roundabout on Rte. 302. At our next stop, Honey Brook, we can make a detailed examination of Pleistocene deposits because the brook has incised the terrace material, exposing extensive sections of nearly all of the valley fill.

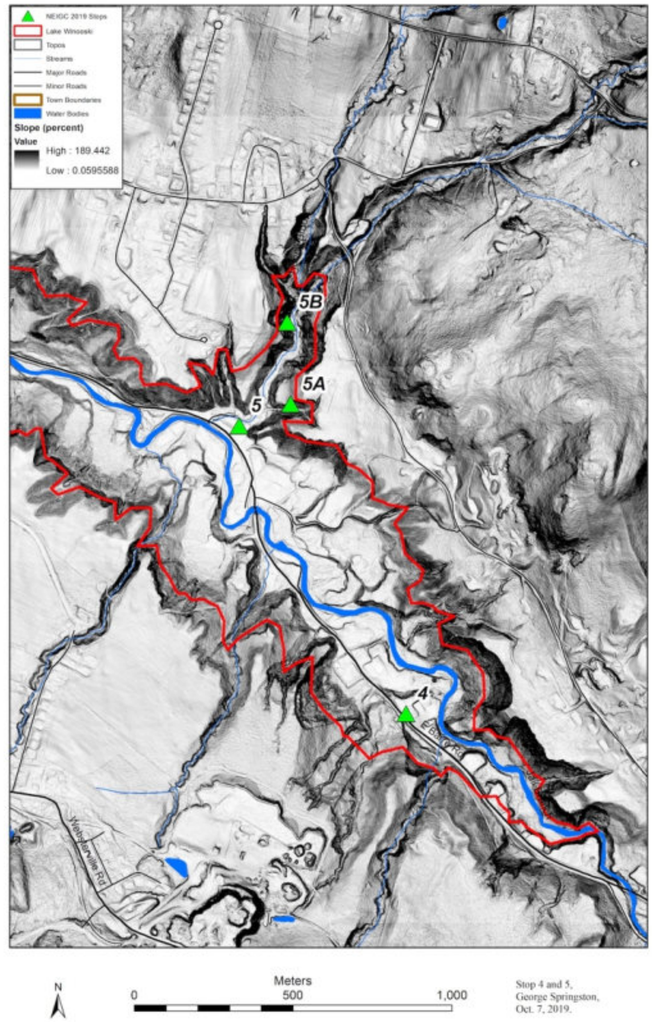


Figure 39: Percent slope map of middle Jail Branch and Honey Brook, a south-flowing tributary. The Jail Branch bedrock valley is partially filled with glaciolacustrine deposits that are capped by a thick diamicton whose surface is fairly flat and ramps gently up the valley, culminating southeast of the Rte. 302 roundabout. Post-glacial incision by Jail Branch has created a wide paired terrace on either side of the valley. From Dunn et al. (2018).

Previously, using outcrops we see at Stop 4 and elsewhere in the Winooski basin, Larsen (1999a; 1999b) and Wright (1999a; 1999b) mapped the glaciolacustrine deposits in the Jail Branch basin as having originated in a pre-glacial lake that Larsen (1999a) called preglacial Lake Merwin. This interpretation was based upon the fact that a till lies upon the lake deposits, that till on bedrock was not present in the outcrops being observed, and that at that time the Middlesex Readvance had not been well established. We reinterpret the sections here in Jail Branch as having been deposited in glacial Lake Winooski and the subsequent glacial Lake East Barre, and the upper till as being from the Middlesex Readvance.

Stop C-5: Honey Brook basin, tributary to Jail Branch. Glacial Lake East Barre (44.1787°N, 72.4698°W, parking area; 44.1792°N, 72.4682°W)

General geologic history preserved in Honey Brook basin

Honey Brook flows south into Jail Branch, with the confluence near the parking location for this stop (Fig. 35). The brook has incised a narrow and steep, canyon-like valley into the Jail Branch terrace. Deposits exposed in Honey Brook were formed in glacial Lake East Barre and are dominated by ice-proximal sediments that demonstrate the advancing of the ice margin and eventual overrunning of the location (Dunn et al., 2018; Dunn et al., 2015; Dunn et al., 2019; Fig. 41). Sedimentation rates were high as high-density turbidity currents and other sediment gravity flows carried sediment from the ice margin to the nearby lake bottom (Fig. 41, Images A, B). The advancing glacier base impacted the lake bottom, producing deformation and also remobilizing large volumes of material that moved as both noncohesive and cohesive masses. Channelized gravel deposits suggest that the lake may have shoaled just before the glacier advanced over this position and folding and faulting of the gravel deposits resulted from glacial tectonism (Fig. 41, Image C).

In both Jail Branch and in Great Brook, which we passed through earlier, the glaciolacustrine section is capped by a massive diamicton, up to 10 m thick, or locally a combination of a thick stratified diamicton overlain by the massive diamicton (Fig. 42). During readvance, the glacier overrode glaciolacustrine sections, producing the massive till, but locally, the glacier seems to have temporarily stalled, perhaps pinned on nearby high topographic points or on the shoaling lakebed, and a thick sequence of a thinly stratified diamicton was deposited before advance recommenced and deposited the massive till. Both will be observed at Stops 5c and 5d.

The stratified diamicton is a major unit locally, and interstratified debrites and glaciolacustrine sands and silts are not particularly unique and can represent an ice cliff margin (Ashley, 1995) where they are attributed to mass-movement, debris slides, and debris squeezed from the base. Thick stratified diamict within proglacial sequences in the Adirondack Uplands, just west of the Champlain Valley, are considered to be from an ice-proximal setting (Franzi et al., 2007). Unlike deposits in our study area, diamict beds in the Adirondacks include lenticular geometry and are interbedded with thin, discontinuous sand, silt and clay

Figure 40: Basal and upper diamictons in Honey Brook, at Stop 5d. Both have a silt matrix and are very compact and both are interpreted as lodgement till. The basal diamicton lies upon bedrock and exhibits a SSE clast fabric while the upper diamicton caps the stratigraphic sequence and has a SSW clast fabric. Difference in color of the upper diamicton is fresh (left) verse weathered (right) and amount of moisture and daylight at time of photographs. If stream conditions are favorable the basal till will be exposed at Stop 5d. It has also been observed immediately across the Jail Branch valley from Honey Brook, where silty rhythmites drape onto the basal till and these are in turn overlain by proximal lake deposits and then a second till. Photos R. Dunn and G. Springston.

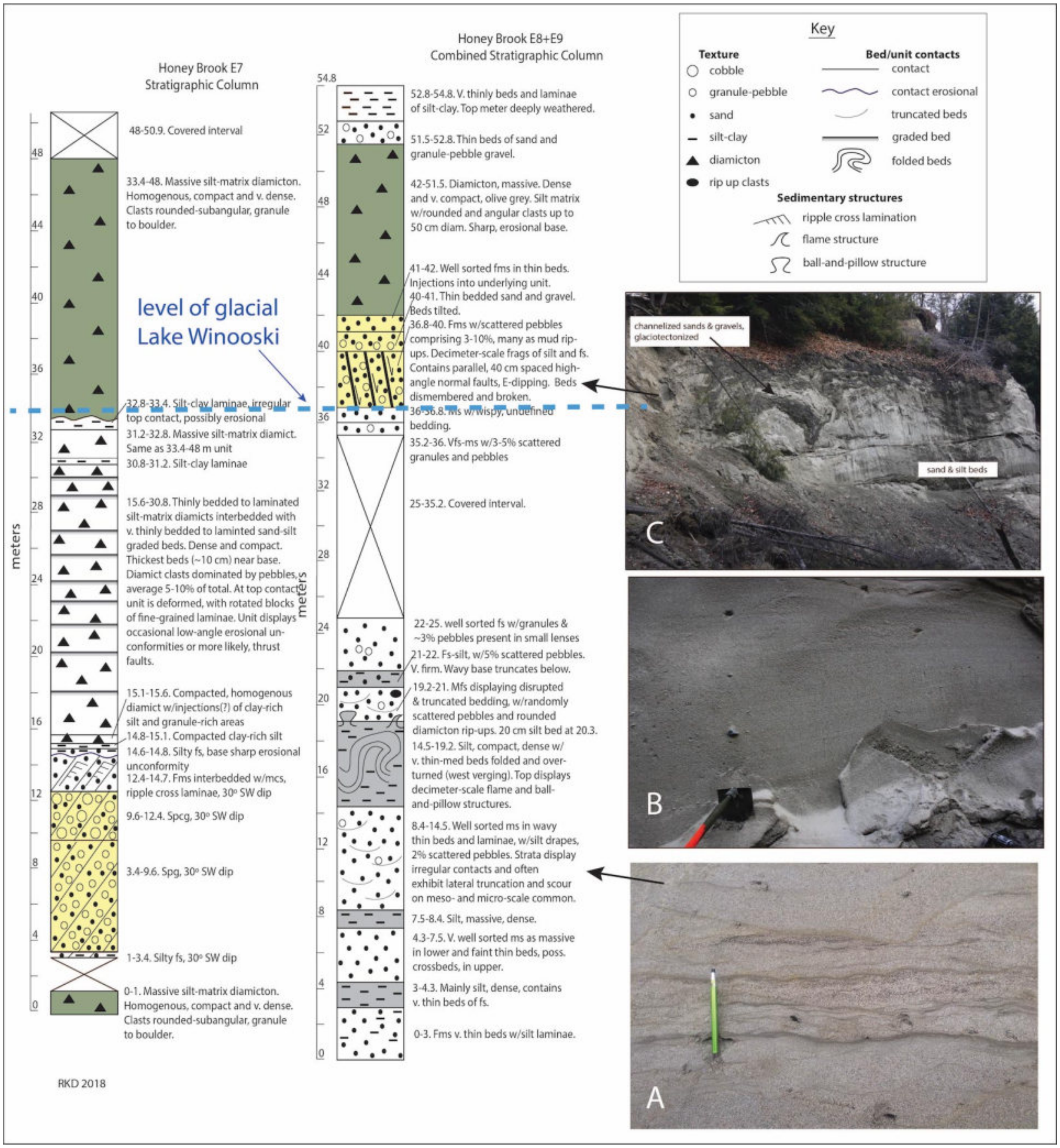


Figure 41: Honey Brook stratigraphic sequence from Stop C-5d (left) and a composite sequence from exposures in the lower valley (right), and images of deposits. **Photo A** demonstrates a common style of deposition in the basin, with bedded sands and silts exhibiting ubiquitous reactivation surfaces and truncation. Beds rarely are traced more than a few meters, depositional lobes seem to be common, grading occurs, dewatering structures can be found, and ripples are rare. Deposition was by sediment gravity flows, including surge- and quasi-steady turbidity flow (Mulder and Alexander, 2001), and grain flow. Likely, these represent remobilized materials originating from glacier impact on the lakebed. **Photo B** shows laminated and homogenous coarse sand with scattered pebbles generated from rapid settling of quasi-steady high-density turbidity flow. **Photo C** shows glaciotectonized sand and gravel that was deposited in channel bodies in the shoaling(?) lake before being overrun by the ice margin. Lodgement till occurs above the exposure but is covered here. Illustrations and photos by R. Dunn

laminae to thick, massive, planar bedded and cross-stratified sand and gravel. Weddle (1992) described interbedded diamict (both stratified and massive) and bedded sands and gravels from a former ice-marginal lake in the Sandy River valley of western Maine. These were interpreted as ice-proximal deposits generated by current driven processes, mass-movement processes, ice-rafting, glaciofluvial processes and subglacial processes (Weddle, 1992). Similar to deposits of this fieldtrip, ice-proximal deposits in the Sandy River valley are partially interbedded with and generally overlie fine-grained distal glaciolacustrine deposits. However, unlike all of these examples, the stratified diamicton in the study area is not interstratified with glaciolacustrine fine sediments and the homogeneity of diamicton production makes the deposits here unique.

Figure 42: Great Brook exposure of stratified diamicton under massive diamicton, both several meters thick. The package is also present locally in Honey Brook. Geologists for scale. Photo G. Springston.

We interpret the stratified diamicton as a package of thin debrites produced at the stalled ice margin as dilatant till was squeezed from the ice base (Dunn et al., 2015, 2018, 2019). A similar mechanism forms grounding zone wedges (GZW) on modern glaciated continental shelves, and they are also recorded in Pleistocene deposits from Whidbey Island, Washington (Anderson et al., 2017; Demet et al., 2017; Simkins et al., 2017; Simkins et al., 2018).

GZW Recent and Pleistocene examples.

Multibeam sonar bathymetry and sub-bottom acoustic studies performed on the continental shelf of Antarctica (Anderson et al., 2017; Simkins et al., 2018) and outcrop studies on Whidbey Island in Washington (Demet et al., 2017; Simkins et al., 2017) have documented very similar deposits to those discussed here. They are similar in size, shape, and composition and are believed to have originated as grounding zone wedges (GZW) during overall glacial ice retreat. GZWs are characterized by a gently sloping stoss face and a relatively steep lee face (3°-34° overall, 4°-10° on Whidbey Is.). Dimensional data from 1,689 grounding zone wedges on the Ross Sea shelf reveal that they average ~6 m in height with a large standard deviation, they average 522 m in width but the mode is 200 m, they are typically asymmetrical in plan form, and they are spaced hundreds of meters apart with large standard deviation (Simkins et al., 2018). Whidbey Island GZWs are similar in size and shape, ranging from approximately 500–1400 m in length and approximately 10-50 m in height (Simkins et al., 2017). On the Ross Sea shelf they were observed under a wide variety of conditions, including variable bed slope and variable water depth, and their construction is encouraged by local pinning points of antecedent topography.

Grounding zone wedges are depositional products in which a basal ice conveyor belt of deforming, dilatant, subglacial till is delivered to the grounding line where it is subsequently transported basinward, developing a wedge-like body with topset aggradation and progradation of foresets (Anderson et al., 2017; Simkins et al., 2017, 2018). Study of the Whidbey Island GZW shows that prograded foresets are composed of diamicts produced by debris flows. Diamicts are commonly interbedded with thin beds of finely laminated sand and mud derived from turbidity flows and sediment-laden meltwater plumes (Demet et al., 2017; Simkins et al., 2017). Study of Ross Sea shelf grounding lines found no evidence that grounding zone wedge production is determined by the presence or absence of an ice shelf (Simkins et al., 2018). Rather, their construction is favored by a high sediment input and a relatively longer-held grounding position. GZWs are generally believed to represent ice margin stability on the order of decades to centuries (Simkins et al., 2018).

While their formation is associated with an overall ice retreat, lineation is often present on Ross Sea grounding zone wedges, indicating readvance over them (Anderson et al., 2017; Greenwood et al., 2018; Simkins et al., 2017; Simkins et al., 2018). In these cases, a retreating ice margin becomes temporarily grounded, sometimes but not always on a topographic high, and this stabilizes the grounding line and results in GZW production. Sediment added to the wedge body decreases the water depth relative to ice thickness, thus overcoming buoyancy effects and allowing the position of grounding to advance, thus producing glacial lineations on the GZW surface (Simkins et al. 2017, 2018).

On Whidbey Island, GZWs are bounded by glacial unconformities (Demet et al. 2017; Simkins et al. 2017). Each GZW is overlain by subglacial till indicating short-term readvance which was a consequence of grounding line stabilization by the same buoyancy reduction mechanism as on the Ross Sea shelf, described above. While readvance over Whidbey Island GZWs was local, with till being deposited on GZWs but not between them, Greenwood et al. (2018) documented the formation of grounding zone wedges and glacial lineation during a Ross Sea ice shelf LGM readvance of at least 50 km.

The ice in upper glacial Lake Winooski was grounded and in addition to deforming lake bottom deposits we believe it locally produced grounding zone bodies similar to those documented elsewhere (Dunn and Wright, 2021). While these bodies may not be a GZW in the strictest sense of the term, they seem to have formed under very similar conditions and for lack of a better term we will call them GZW for this report.

Stop C-5a. Optional

A short distance upstream from the parking area a small side 'canyon' to the east contains abundant exposures of materials from glacial Lake East Barre. The various exposures here are dominated by ice-proximal glaciolacustrine deposits (Dunn et al., 2015; Fig. 41, composite section). Moving up section one can observe many examples of sediment gravity flows, and noncohesive and cohesive deposits from remobilized lake bottom material (Fig. 41, Images A, B). A 1+ meter thick silt body near the top of the section includes extensive internal deformation, and the upper surface includes a very large flame structure that is many decimeters high and wide. This plastic-behaving silt was remobilized as a cohesive package, but with internal deformation, presumably by glaciotectonism of the lakebed. We will see similar silt "slides" at Stops 5b and 5c. Most notable in the upper part of the total sequence are: 1. large channel bodies of sandy gravel at about the level of glacial Lake Winooski; 2. glaciotectonism of the channel gravels and other outwash (Fig. 41, Image C); and 3. a thick, dense till above glaciolacustrine deposits. In this area, the lake was filling rapidly to the point that it shoaled and then channels were incised by gravel-rich meltwater. The gravel-filled channels were then immediately overrun by the advancing ice.

Stop C-5b.

Here, in the lower Honey Brook basin, a section displays various types of remobilized deposits that were dislodged by glaciotectonism of the lakebed, and possibly by glacioseismicity. Materials moved as semi-cohesive and cohesive masses, including sediment gravity flows and slides that sometimes contain extensive internal deformation (Fig. 43). Their rapid emplacement often produced dewatering structures in underlying silty sands.

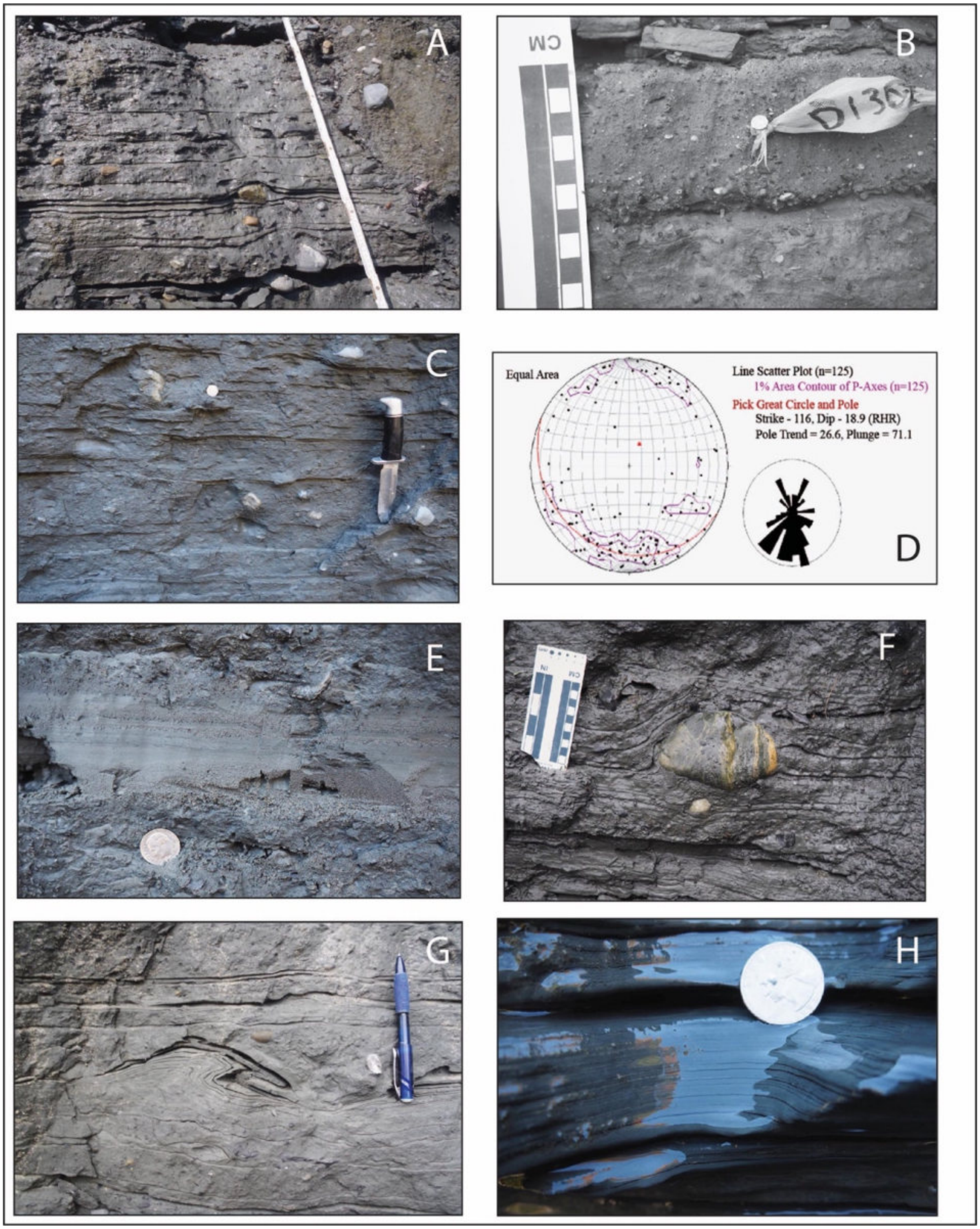


Figure 44: Examples of sedimentary features of the stratified diamicton. **10-A**. Photo from Great Brook showing thin debrite beds alternating with graded very thin beds and laminations (fine sand to silt). Sand in the graded beds washes out revealing the stratification. **10-B**. Normal graded bed in a debrite, reverse grading is also present. **10-C.** Stratified diamicton at Stop 5d in Honey Brook. **10-D.** Fabric analysis of the Honey Brook stratified diamicton. **10-E.** Several thin graded beds which are typically found between debrites. **10-F**. Dropstone in stratified diamicton at Great Brook. **10-G.** Drag fold in several graded laminae, probably produced during emplacement of the overlying diamicton (behind pencil). **10-H.** Several graded laminae. Debrites are almost always separated by one or more graded beds or laminae, probably representing a hybrid flow mechanism (Haughton et al., 2009). Photos R. Dunn.

Stop C-5c. High mass failure on west side of mid-Honey Brook valley
Prior to the July 2023 flash flood this location was a small mass failure, but now presents nearly a complete vertical sequence of the valley fill. The lower half of the section comprises glacioloacustrine deposits that coarsen upward, from mostly silty sands at the base to gravels in the mid-slope. A couple of silt horizons in the lower part of the section are interpreted as remobilized slide material. They exhibit erosional truncation of underlying sediments and injection into the same. Calcium carbonate concretions are common in Honey Brook deposits and are found throughout this silt horizon. Overlying the gravels is ~8 m of stratified diamicton that is in turn overlain by ~16 m of massive diamicton, the top of which is covered. The sequence records the advancing ice margin, its stalling and deposition of a grounding zone body, and subsequent overriding of the site and deposition of lodgement till.

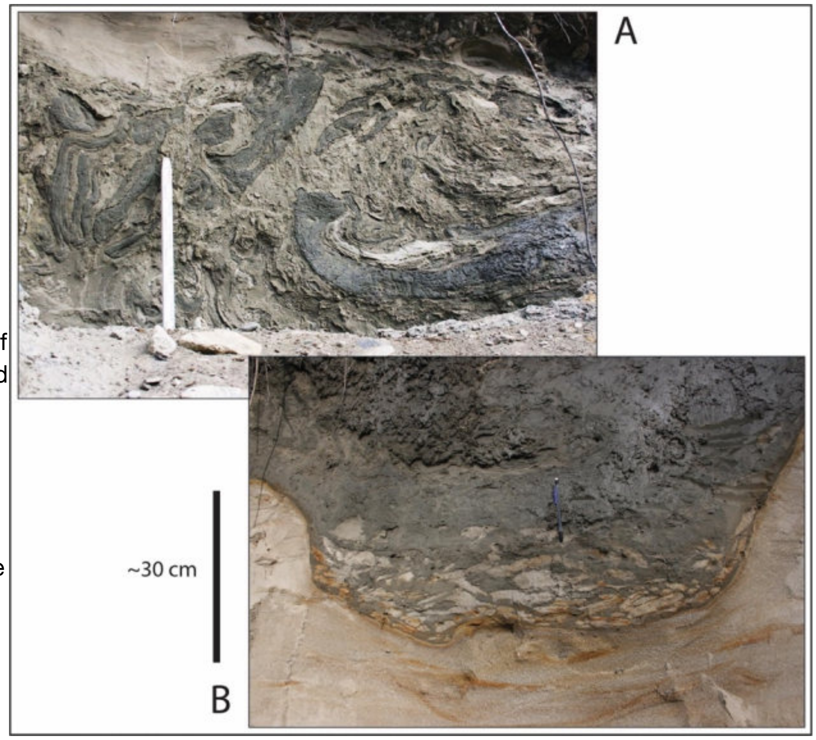


Figure 43: Semi-cohesive and cohesive subaqueous mass failure deposits at Stop 5b. **A** is a slide ~2 m thick consisting of interbedded sand and silt with an abundance of internal deformation demonstrating plastic and brittle behavior. Folds are complex and dismembered, and some limbs are fractured. The slide top is irregular, and the base is an erosional surface but in places is injected into underlying sand. Stick one meter in length. **B** is a close-up of a small portion of the base of a silt-rich slide deposit. Most of the base is an erosion surface but locally, as here, it displays scours in channel-form that include rip-ups of underlying sand beds. Intact internal bedding reveals that the slide moved cohesively, but bedding at the base is dismembered, chaotically mixed, or homogenized, possibly indicating fluid entrainment from below. The upper surfaces of slides often display large-scale dewatering structures. Photos R. Dunn.

Stop C-5d. High mass failure on west side of upper Honey Brook valley; last exposure before granite falls
This ~50 m high exposure includes a basal lodgement till that can be seen if stream conditions are favorable. The till is overlain by bedded sand and sandy gravel that dip steeply to the south, down the valley (Fig. 47, left column). The sand-gravel package correlates to the same units at Stop C-5c. Above the gravel, and lying upon it unconformably, is approximately 16 m of stratified diamicton consisting of thin bedded debrites that are typically separated by very thin graded beds or thin-medium silt beds (Fig. 44). The stratified diamicton has several attributes similar to ground zone wedges (isolated occurrence, size, stratigraphic position, depositional mechanism), and is thus interpreted as a GZW-like body, as is the same unit in Great Brook (Dunn et al., 2018, 2019; Dunn and Wright, 2021; Fig. 42). The vertical sequence here is capped by ~15 meters of massive, compact diamicton interpreted as lodgment till.

Interestingly, as noted each debrite bed is typically associated with an overlying or underlying, usually very thin, graded bed(s), and while we believe they are genetically related, we have yet to determine if the over or under graded bed is the pair with a given debrite (Fig. 44). The answer to this lies in the type of flow that occurred; for example, a debris flow moving with semi-plasticity may undergo reverse shearing at the water–sediment boundary, creating an overriding turbid current(s) that eventually creates a graded bed(s) above the diamict of the debris flow. Conversely, some hybrid flows consist of a turbidity current that runs in front of a debris flow, resulting in a genetically related graded bed below a diamict (Haughton et al., 2009; Talling, 2013; Talling et al., 2012). Whatever the exact mechanism, the thickness of the overall stratified diamict body and the highly repetitive nature of deposition indicates a high volume, local sediment source and a consistent sediment delivery mechanism.

Overall, the lower portion of the GZW section exposed at Honey Brook contains more silt and silty-clay beds than the GZW in Great Brook and perhaps represents a distal portion of a prograding GZW. A probable thrust fault can be found near the base of the stratified diamicton, but it can be difficult to find depending upon conditions on the exposure. A wide, low-relief channel body filled with laminated silts and thin graded beds is also present and probably represents a higher discharge rate at this point (Demet et al., 2017; Dunn and Wright, 2021).

Sunday Summary

In summary, unlike the well sorted, bedded sands and rhythmites common in the main glacial Lake Winooski basin, sections at Nasmith Brook, Great Brook, and Jail Branch all comprise a thick till over ice-proximal glaciolacustrine sediments that typically reveal conditions of rapid sedimentation, remobilization, and deformation because of the approaching and overriding of a glacier margin. Locally, thick stratified diamicton was deposited at the grounding line, presumably when the ice temporarily stalled. In upper glacial Lake Winooski, readvance of the ice margin over an unknown distance produced small, isolated proglacial lakes whose sedimentary record captures the dynamic conditions of the subaqueous ice-marginal environment.

References

Ackerly, S. C., 1989, Reconstructions of mountain glacier profiles, northeastern United States: Geological Society of America Bulletin, v. 101, p. 561-572.

Ackerly, S. C., and Larsen, F. D., 1987, Southwest-trending striations in the Green Mountains, central Vermont, *in* Westerman, D. S., ed., New England Intercollegiate Geological Conference Guidebook, Volume 79: Northfield, Vermont, p. 369–382.

Anderson, J. B., Simkins, L., Greenwood, S. L., Demet, B., and Halberstadt, A. R., 2017, Role of grounding zone wedges in grounding line stabilization: case study from Ross Sea: Geological Society of America Abstracts with Programs, v. 49.

Antevs, E., 1922, The Recession of the Last Ice Sheet in New England with a preface and contributions by J. W. Goldthwait, New York,, American Geographical Society, American Geographical Society research series, v. no 11, 4 p. l., vii -xiii, 120 p. p.:

-, 1928, The last glaciation : with special reference to the ice retreat in northeastern North America, New York, American geographical society, American Geographical Society research series, v. no 17, x, 292 p. p.:

Ashley, G. M., 1995, Glaciolacustrine environments, *in* Menzies, J., ed., Modern Glacial Environments; Processes, dynamics and sediments: Oxford, Butterworth and Heinemann, p. 417-444.

Balco, G., Stone, J. O. H., Porter, S. C., and Caffee, M. W., 2002, Cosmogenic-nuclide ages for New England coastal moraines, Martha's Vineyard and Cape Cod, Massachusetts, USA: Quaternary Science Reviews, v. 21, p. 2127-2135.

Barg, L., and Springston, G. E., 2001, Assessment of fluvial geomorphology in relation to hazards from riverine erosion and landslides in the Great Brook watershed in Plainfield, Vermont: Manuscript submitted to the Vermont Geological Survey, p. 1-71.

Benn, D. I., and Hulton, N. R. J., 2010, An ExcelTM spreadsheet program for reconstructing the surface profile of former mountain glaciers and ice caps: Computers & Geosciences, v. 36, p. 605-610.

Bierman, P. R., Davis, P. T., Corbett, L. B., Lifton, N. A., and Finkel, R. C., 2015, Cold-based Laurentide ice covered New England's highest summits during the Last Glacial Maximum: Geology, v. 43, p. 1059-1062.

Bierman, P. R., Lini, A., Davis, P. T., Southon, J., Baldwin, L., Church, A., and Zehfuss, P. H., 1997, Post-glacial ponds and alluvial fans: Recorders of Holocene landscape history: GSA Today, v. 7, p. 1-8.

Chapman, D. H., 1937, Late-glacial and postglacial history of the Champlain Valley: American Journal of Science, v. Series 5 Vol. 34, no. 200, p. 89-124.

Connally, G. G., 1972, Proglacial lakes in the Lamoille valley, Vermont, *in* Doolan, B. L., and Stanley, R. S., eds., New England Intercollegiate Geological Conference Guidebook, Volume 64, p. 343–358.

Connally, G. G., and Calkin, P. E., 1972, Woodfordian glacial history of the Champlain lowland, Burlington to Brandon, Vermont, *in* Doolan, B. L., and Stanley, R. S., eds., New England Intercollegiate Geological Conference, Volume 64, p. 389-397.

Corbett, L. B., Bierman, P. R., Stone, B. D., Caffee, M. W., and Larsen, P. L., 2017, Cosmogenic nuclide age estimate for Laurentide Ice Sheet recession from the terminal moraine, New Jersey, USA, and constraints on latest Pleistocene ice sheet history: Quaternary Research, v. 87, no. 3, p. 482-498.

Corbett, L. B., Bierman, P. R., Wright, S. F., Shakun, J. D., Davis, P. T., Goehring, B. M., Halsted, C. T., Koester, A. J., Caffee, M. W., and Zimmerman, S. R., 2019, Analysis of multiple cosmogenic nuclides constrains Laurentide Ice Sheet history and process on Mt. Mansfield, Vermont's highest peak: Quaternary Science Reviews, v. 205, p. 234-246.

Davis, P. T., 1999, Cirques of the Presidential Range, New Hampshire, and surrounding alpine areas in the northeastern United States: Géographie physique et Quaternaire, v. 53, p. 25-45.

-, 2024, Till benches mark receding late Wisconsinan ice margins in mountainous terrain: A case study from Mount Moosilauke, NH: Geological Society of America Abstracts with Programs, v. 56.

Davis, P. T., Keeley, J. A., and Thompson, P. J., 2022, Surficial Geologic Map of the Mount Moosilauke 7.5' Quadrangle, New Hampshire: New Hampshire Surficial Geologic Map Open-File Series, v. GEO-061-024000-SMOF.

Demet, B. P., Nittrouer, J. A., Anderson, J. B., and Simkins, L. M., 2017, Sedimentary processes at ice sheet grounding-zone wedges: examples from Antarctica and Washington State (U.S.A.): Geological Society of America Abstracts with Programs, v. 49.

Dunn, R. K., Morin, J., and Hermanson, T., 2018, A proglacial ice margin depositional sequence marked by sedimentation in a grounding zone wedge and by remobilization of basin slopes: Geological Society of America Abstracts with Programs, v. 50.

Dunn, R. K., and Springston, G. E., 2019, A lacustrine record of Late Pleistocene ice readvance in the Winooski basin of central Vermont, *in* Koteas, C., ed., New England Intercollegiate Geological Field Conference, Volume 111, p. 23-45.

Dunn, R. K., Springston, G. E., and Donahue, N. P., 2007, Surficial Geologic Map of the Mad River Watershed, Vermont (northern and southern sheets): Vermont Geological Survey Open-File Report VG07-1a,b, scale 1:24,000.

Dunn, R. K., Springston, G. E., Hermanson, T., and Thomas, E. K., 2015, Interbedded subaqueous debris and turbidity flows; a thick and laterally extensive ice-proximal facies preserved in isolated proglacial basins: Geological Society of America Abstracts with Programs, v. 47, p. 113.

Dunn, R. K., Springston, G. E., Rogers, K., and Loeffler, A., 2019, Lacustrine deposition in the ice-proximal environment during glacial advance, a model from central Vermont: Geological Society of America Abstracts with Programs, v. 51.

Dunn, R. K., Springston, G. E., and Wright, S. F., 2011, Quaternary Geology of the Central Winsooski River Watershed with focus on Glacial Lake History of Tributary Valleys (Thatcher Brook and Mad River), *in* West, D. P., Jr., ed., New England Intercollegiate Geological Conference Guidebook, Volume 103, p. C3-1-C3-32.

Dunn, R. K., and Wright, S. F., 2021, Grounding zone deposits associated with streaming readvance of the Laurentide ice sheet in the Winooski valley of northern Vermont: Geological Society of America Abstracts with Programs, v. 53.

Franzi, D. A., Rayburn, J. A., Knuepfer, P. L., and Cronin, T. M., 2007, Late Quaternary history of northeastern New York and adjacent parts of Vermont and Quebec: Guidebook for the 70th Annual Meeting of the Northeast Friends of the Pleistocene, Plattsburgh, New York, v. 70, p. 1-70.

Geier, G., 2021, The surficial geology of Mount Moosilauke, New Hampshire, and an investigation of slope-parallel ridges: A LiDAR approach: Dartmouth College Senior Thesis.

Greenwood, S. L., Simkins, L. M., Halberstadt, A. R., Prothro, L. O., and Anderson, J. B., 2018, Holocene reconfiguration and readvance of the East Antarctic Ice Sheet: Nature Communications, v. 9.

Grigg, L. D., Engle, K. J., Smith, A. J., Shuman, B. N., and Mandl, M. B., 2021, A multi-proxy reconstruction of climate during the late-Pleistocene to early Holocene transition in the northeastern USA: Quaternary Research, v. 114, p. 1-17.

Grigg, L. D., Stefanescu, I., Shuman, B. N., and Oswald, W. W., 2023, A pollen-inferred reconstruction of Holocene climatic and ecological variability from Vermont, USA: Quaternary Research, v. 116, p. 60-77.

Halsted, C. T., Bierman, P. R., Shakun, J. D., Davis, P. T., Corbett, L. B., Caffee, M. W., Hodgdon, T. S., and Licciardi, J. M., 2022, Rapid southeastern Laurentide Ice Sheet thinning during the last deglaciation revealed by elevation profiles of in situ cosmogenic 10Be: Geological Society of America Bulletin.

Haughton, P., Davis, C., McCaffrey, W., and Barker, S., 2009, Hybrid sediment gravity flow deposits – Classification, origin and significance: Marine and Petroleum Geology, v. 26, p. 1900-1918.

Jennings, K. L., Bierman, P. R., and Southon, J., 2003, Timing and style of deposition on humid-temperate fans, Vermont, United States: Geological Society of America Bulletin, v. 115, no. 2, p. 182-199.

Koteff, C., and Larsen, F. D., 1989, Postglacial uplift in western New England: Geologic evidence for delayed rebound, *in* Gregerson, S., and Basham, P. W., eds., Earthquakes at North Atlantic Passive Margins: Neotectonics and Postglacial Rebound, Kluwer, Norwell, p. 105-123.

Larsen, F. D., 1972, Glacial History of Central Vermont, *in* Doolan, B. L., and Stanley, R. S., eds., New England Intercollegiate Geological Conference Guidebook, Volume 64, p. 296-316.

-, 1987, History of Glacial Lakes in the Dog River Valley, Central Vermont, *in* Westerman, D. S., ed., New England Intercollegiate Geological Conference Guidebook, Volume 79, p. 213-236.

Larsen, F. D., 1999a, Glacial history of the Montpelier, Vermont, 7.5-minute Quadrangle, *in* Wright, S. F., ed., New England Intercollegiate Geological Field Conference Guiidebook, Volume 91, p. 286-300.

Larsen, F. D., 1999b, Surficial Geologic Map of the Montpelier Quadrangle, Vermont: Vermont Geological Survey, scale 1:24,000.

Larsen, F. D., 2001, The Middlesex Readvance of the Late-Wisconsinan ice sheet in central Vermont at 11,900 C-14 years BP: Geological Society of America Abstracts with Programs, v. 33, no. 1.

Larsen, F. D., Dunn, R. K., Donahue, N. P., and Springston, G. E., 2003a, Surficial Geology of the Watershed of the Third Branch of the White River, Central Vermont: Unpublished Open File Map, Vermont Geological Survey.

Larsen, F. D., Ridge, J. C., and Wright, S. F., 2001, Correlation of varves of Glacial Lake Winooski, north-central Vermont: Geological Society of America Abstracts with Programs, v. 33, no. 1.

Larsen, F. D., Wright, S. F., Springston, G. E., and Dunn, R. K., 2003b, Glacial, Late-Glacial, and Postglacial History of Central Vermont, Volume 66, p. 62.

Loso, M. G., Schwartz, H. K., Wright, S. F., and Bierman, P. R., 1998, Composition, morphology, and genesis of a moraine-like feature in the Miller Brook Valley, Vermont: Northeastern Geology and Environmental Sciences, v. 20, p. 1-10.

MacClintock, P., 1971, Pleistocene mountain glaciation, Northern Vermont: Discussion: Geological Society of America Bulletin, v. 82, p. 1760-1761.

Magdon, I., Dutil, M., and Grigg, L. D., 2018, A Late-glacial record of environmental change from Pecks Pond wetland, VT: Geological Society of America Abstracts with Programs, v. 50, no. 2.

Mandl, M. B., Shuman, B. N., Marsicek, J., and Grigg, L. D., 2016, Estimating the regional climate signal in a late Pleistocene and early Holocene lake-sediment δ 18O record from Vermont, USA: Quaternary Research, v. 86, p. 67-78.

Merwin, H. E., 1908, Some late Wisconsin and Post-Wisconsin Shore-lines of North-western Vermont, Vermont State Geologist 6th Report, Volume 6, p. 113--138.

Munroe, J. S., Perzan, Z. M., and Amidon, W. H., 2016, Cave sediments constrain the lastest Pleistocene advance of the Laurentide Ice Sheet in the Champlain Valley, Vermont, USA: Journal of Quaternary Science, v. 31, no. 8, p. 893-904.

Noren, A. J., Bierman, P. R., Steig, E. J., Lini, A., and Southon, J., 2002, Millennial-scale storminess variability in the northeastern United States during the Holocene epoch: Nature, v. 419, no. 6909, p. 821-824.

Oswald, W. W., and Foster, D. R., 2011, Middle-Holocene dynamics of Tsuga canadensis (eastern hemlock) in northern New England, USA: The Holocene, v. 22, p. 71-78.

Oswald, W. W., Foster, D. R., Shuman, B. N., Doughty, E. D., Daison, E. K., Hall, B. R., Hansen, B. C., Lindbladh, M., Marroquin, A., and Truebe, S. A., 2018, Subregional variability in the response of New England vegetation to postglacial climate change: Journal of Biogeography, v. 45, p. 2375-2388.

Parent, M., and Occhietti, S., 1999, Late Wisconsinan deglaciation and glacial lake development in the Appalachians of southeastern Québec: Géographie physique et Quaternaire, v. 53, no. 1, p. 117-135.

Ratcliffe, N. M., Walsh, G. J., Gale, M. H., Masonic, L. M., Estabrook, J. R., Geological Survey (U.S.), Vermont. Agency of Natural Resources., and Vermont Geological Survey., 2011, Bedrock geologic map of Vermont, scale 1:100,000.

Rayburn, J. A., Cronin, T. M., Franzi, D. A., Knuepfer, P. L. K., and Willard, D. A., 2011, Timing and duration of North American glacial lake discharges and the Younger Dryas climate reversal: Quaternary Research, v. 75, no. 3, p. 541-551.

Rayburn, J. A., Knuepfer, P. L., and Franzi, D. A., 2005, A series of large, Late Wisconsinan meltwater floods through the Champlain and Hudson Valleys, New York State, USA: Quaternary Science Reviews, v. 24, no. 22, p. 2410-2419.

Ridge, J. C., 2016, The North American Varve Project, https://eos.tufts.edu/varves/images/navc_deglac_NEiceMargmapsINTCAL13-August2016_big.jpg.

Ridge, J. C., Balco, G., Bayless, R. L., Beck, C. C., Carter, L. B., Dean, J. L., Voytek, E. B., and Wei, J. H., 2012, The new North American Varve Chronology: A precise record of southeastern Laurentide Ice Sheet deglaciation and climate, 18.2-12.5 kyr BP, and correlations with Greenland ice core records: American Journal of Science, v. 312, no. 7, p. 685-722.

Simkins, L. M., Anderson, J. B., and Demet, B. P., 2017, Grounding line processes of the southern Cordilleran Ice Sheet in Puget Lowland, *in* Haugerud, R. A., and Kelsey, H. M., eds., From the Puget Lowland East of the Cascade Range: Geological Excursions in the Pacific Northwest: Geological Society of America Field Guide, Volume 49, p. 53-65.

Simkins, L. M., Greenwood, S. L., and Anderson, J. B., 2018, Diagnosing ice sheet grounding line stability from landform morphology: The Cryosphere, v. 12, p. 2702-2726.

Sperling, J. A., Wehrle, M. E., and Newman, W. S., 1989, Mountain glaciation at Ritterbush Pond and Miller Brook, northern Vermont, Reexamined: Northeastern Geology and Environmental Sciences, v. 11, p. 106-111.

Springston, G. E., 2011, Surficial Geologic Map of the Plainfield Quadrangle, Vermont: Vermont Geological Survey Open-File Report VG11-4, scale 1:24,000.

-, 2017, Landslide inventory of Washington County, central Vermont: Vermont Geological Survey Open File Report VG2017-7.

-, 2018, Surficial geology of the Barre East 7 1/2 minute quadrangle, Vermont: Vermont Geological Survey Open File Report VG2018-1.

Springston, G. E., and Barg, L., 2002, Surficial geology of the Great Brook watershed, central Vermont: Manuscript submitted to the Vermont Geological Survey.

Springston, G. E., and Dunn, R. K., 2006, Surficial Geologic Map of the Southern Worcester Mountains Watershed, Middlesex and Stowe: Vermont Geological Survey Open-File Report VG2006-3, scale 1:24,000.

Springston, G. E., and Kim, J. J., 2008, Surficial Geologic Map of the Knox Mountain Area, Marshfield and Peacham, Vermont: Vermont Geological Survey Open-File Report VG08-4, scale 1:24,000.

Springston, G. E., Kim, J. J., DeJong, B., Boyles, J. L., and Klepeis, K., 2023, Two decades of collaborative landslide research in Vermont: Geological Society of America Abstracts with Programs, v. 55.

Springston, G. E., and Thomas, E. K., 2014, Landslide hazard analysis of the Great Brook watershed, Plainfield, Vermont: Report prepared for the Central Vermont Regional Planning Commission, Montpelier, Vermont, p. 1-34.

-, 2018, Landslides in the Great Brook watershed, Washington County, central Vermont: Geological Society of America Abstracts with Programs, v. 50.

Stewart, D. P., 1971, Pleistocene mountain glaciation, Northern Vermont: Discussion: Geological Society of America Bulletin, v. 82, p. 1759-1760.

Stewart, D. P., and MacClintock, P., 1969, The Surficial Geology and Pleistocene History of Vermont: Montpelier, Vt.,, Dept. of Water Resources, p. 251.

-, 1970, Surficial Geologic Map of Vermont: Vermont Geological Survey, scale 1:250,000.

Talling, P. J., 2013, Hybrid submarine flows comprising turbidity current and cohesive debris flow: Deposits, theoretical and experimental analyses, and generalized models: Geosphere, v. 9, no. 3, p. 460-488.

Talling, P. J., Masson, D. G., Sumner, E. J., and Malgesini, G., 2012, Subaqueous sediment density flows: Depositional processes and deposit types: Sedimentology, v. 59, p. 1937-2003.

Thompson, W. B., and Borns, H. W., 1985, Surficial Geologic Map of Maine: Maine Geological Survey, scale 1:500,000.

Thompson, W. B., Dorion, C. C., Ridge, J. C., Balco, G., Fowler, B. K., and Svendsen, K. M., 2017, Deglaciation and late-glacial climate change in the White Mountains, New Hampshire, USA: Quaternary Research, v. 87, no. 1, p. 96-120.

Thompson, W. B., Fowler, B. K., and Dorion, C. C., 1999, Deglaciation of the northwestern White Mountains, New Hampshire: Geographie Physique et Quaternaire, v. 53, no. 1, p. 59-77.

Van Hoesen, J. G., Springston, G. E., Franzi, D. A., and Wright, S. F., 2016, A cartographic ode to Chapman: A revised regional depiction of postglacial landscape evolution in the Champlain Valley: Geological Society of America Abstracts with Programs, v. 48.

Wagner, W. P., 1970, Pleistocene mountain glaciation, Northern Vermont: Geological Society of America Bulletin, v. 81, p. 2465-2470.

-, 1971, Pleistocene mountain glaciation, Northern Vermont: Reply: Geological Society of America Bulletin, v. 82, p. 1761-1762.

-, 1972, Ice margins and water levels in northwestern Vermont, *in* Doolan, B. L., and Stanley, R. S., eds., New England Intercollegiate Geological Conference Guidebook, Volume 64, p. 317-331.

Waitt, R. B., and Davis, P. T., 1988, No evidence for post-ice sheet cirque glaciation in New England: American Journal of Science, v. 288, p. 495-533.

Weddle, T. K., 1992, Late Wisconsinan stratigraphy in the lower Sandy River valley, New Sharone, Maine: Geological Society of America Bulletin, v. 104, p. 1350-1363.

Wright, S. F., 1999a, Deglaciation history of the Stevens Branch Valley, Williamstown to Barre, Vermont, *in* Wright, S. F., ed., New England Intercollegiate Geological Conference Guidebook, Volume 91, p. 179-199.

Wright, S. F., 1999b, Surficial Geologic Map of the Barre West Quadrangle, Vermont: Vermont Geological Survey Open-File Report VG99-6, scale 1:24,000.

-, 2001, Surficial Geology of the Jeffersonville Quadrangle, Vermont: Vermont Geological Survey Open-File Reports VG01-2, scale 1:24,000.

Wright, S. F., 2015a, Extent of the Middlesex Readvance in the Winooski River basin, Northern Vermont: Geological Society of America Abstracts with Programs, v. 47, no. 3, p. 83.

-, 2015b, Late Wisconsinan ice sheet flow across northern and central Vermont, USA: Quaternary Science Reviews, v. 129, p. 216-228.

-, 2017, Ice streaming in the southern Champlain and northern Hudson River valleys, Vermont and New York: Geological Society of America Abstracts with Programs, v. 49, no. 6.

-, 2018a, The evolution of glacial lakes in the Winooski River Valley, Vermont: Geological Society of America Abstracts with Programs, v. 50, no. 2.

-, 2018b, Surficial Geology and Hydrogeology of the Bolton Mountain Quadrangle, Vermont: Vermont Geological Survey Open File Report VG2018-4, scale 1:24,000.

-, 2019a, Glacial geology of the Miller Brook Valley, Stowe, Vermont: A head to toe traverse, *in* Koteas, C., ed., New England Intercollegiate Geological Conference, Volume 111, p. 109-130.

-, 2019b, Lateral recessional moraines in the Green Mountains of northern Vermont: Geological Society of America Abstracts with Programs, v. 51, no. 1.

-, 2020, Surficial geology and groundwater hydrology of the Stowe 7.5 minute quadrangle, Vermont: Vermont Geological Survey Open File Report VG2020-1, Report plus 5 maps.

-, 2022a, Intertwined histories of Glacial Lake Hitchcock and Glacial Lake Winooski in the Brookfield Quadrangle, north-central Vermont: Geological Society of America Abstracts with Programs, v. 54, no. 3.

-, 2022b, Surficial Geology and Groundwater Hydrology of the Brookfield 7.5-minute Quadrangle, Vermont: Vermont Geological Survey Open File Report VG2022-1, Report plus 5 maps.

-, 2022c, Surficial Geology and Groundwater Hydrology of the Lincoln 7.5-minute Quadrangle, Vermont: Vermont Geological Survey Open File Report VG2022-3, Report plus 5 maps.

-, 2022d, Surficial Geology and Groundwater Hydrology of the Mount Ellen 7.5-minute Quadrangle, Vermont: Vermont Geological Survey Open File Report VG2022-2, Report plus 5 maps.

-, 2023, Landslides in glacial lakes: A fundamental process governing the stratigraphy of glacial lakes in mountainous terrains: Geological Society of America Abstracts with Programs, v. 55, no. 2.

-, 2024, Ice-marginal channels and the evolution of Glacial Lake Memphremagog along Vermont's northern border: Geological Society of America Abstracts with Programs. Vol. 56, No. 1.

Wright, S. F., Bierman, P. R., Whalen, T. N., and Zehfuss, P. H., 1997, Late Pleistocene-Holocene history; Huntington River and Miller Brook valleys, northern Vermont: Annual Meeting - New England Intercollegiate Geological Conference, v. 89, p. C4.1-C4.30.

Wright, S. F., Dunn, R. K., and Springston, G. E., 2023a, Surficial Geologic Map and Cross-Sections of the Waterbury 7.5-minute Quadrangle, Vermont, v. Vermont Geological Survey Open File Report VG2023-1.

Wright, S. F., Springston, G. E., and Dunn, R. K., 2023b, Surficial Geologic Map and Cross-Sections of the Middlesex 7.5-minute Quadrangle, Vermont, v. Vermont Geological Survey Open File Report VG2023-2.