For GSA field Trip, Boston 2001

Post-glacial Surface Processes of Northern New England

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ABSTRACT

The last 13,000 years of New England's hillslope and fluvial history are preserved in deglacial sediments, alluvial fans, pond sediments, and river terraces. The altitude-dependent trends of glacial striae demonstrate that the mountainous terrain of northwestern Vermont channeled thinning ice during deglaciation, leaving the mountain peaks as nunataks. The Green Mountains hold a variety of evidence that has been interpreted to suggest the presence of alpine glaciers in Vermont after Laurentide ice retreated; recent work argues that alpine glaciers are not necessary to explain these geologic features. Glacial lakes were a dominant feature of the deglacial landscape; glacial-lacustrine and glacial-marine sediments are well exposed by landslides. Tundra vegetation at the top of Vermont's highest peaks, including Mt. Mansfield, is a relict from times when the climate throughout the region was much cooler. The history of Holocene hillslope activity is preserved in pond sediments and revealed by trenches cut into fans and terraces, deposited throughout the Holocene, including the present. Rock falls and debris avalanches are present in the Green Mountains, as well as evidence of the geologic effects of historic clear cutting.

INTRODUCTION

The landscape of northwestern Vermont is one of dramatic contrasts. To the west are the Champlain Lowland and Lake Champlain, draining north to the Gulf of Saint Lawrence over a bedrock-controlled spillway in the Richelieu River (Fig. 1). To the east are the Green Mountains, oriented north-south following the strike of the dominant foliation, the structural grain of the schists and phyllites that dominate the range. The Champlain Lowland, underlain predominately by sedimentary rocks, is mantled in many places by sorted glacial sediments, deposited directly off ice or in glacial lakes bordering the ice margin. The uplands of the Green Mountains, in places where bedrock does not crop out, are covered primarily with varying thickness of till. In the uplands, sorted glacial and post-glacial sediments are rarely present outside river valleys, with the exception of isolated ice-marginal deposits (Stewart and MacClintock, 1969).

Presumably, northern Vermont was repeatedly overrun by advancing icesheets throughout the Quaternary. The latest advance probably overran the state sometime before 27,000 ^{14}C y BP (Fullerton, 1986), at its maxi-

Figure 4: Historic landscape response, Chittenden County, northwestern Vermont. Open circles represent percentage farmland. Major expansion and contraction of Winooski River delta in Champlain as deduced from historic maps are marked with arrows.

mum burying northern Vermont under several kilometers of ice. Deglaciation appears to have occurred by thinning, separation over the mountains, marginal retreat, and eventual stagnation near the ice margin as a calving bay advanced up the St. Lawrence River Valley (Chauvin et al., 1985). The only direct age limit we have for the onset of deglaciation (>12.7 ^{14}C y BP) in northern Vermont comes from a ^{14}C age of bulk organic material at the base of a core from Sterling Pond, 900 m elevation on the flank of Mt. Mansfield (Li, 1996). Basal ^{14}C ages of cores collected from Ritterbush Pond near Eden, Vermont, 600 m lower in elevation and 40 km to the northwest, are 800 ^{14}C years younger (Li, 1996), consistent with a thinning icesheet and northwesterly margin retreat.

Although New England is often referred to as a landscape shaped by ice, it is hard to know exactly the actual impact of glaciers. Without question, icesheets molded the pre-existing landscape, removing most of the weathered rock and shaping the rock outcroppings. The depth of material removed from northern Vermont by glacial action is unknown, but in some places, surprisingly little material was scoured. For example, in the Champlain Lowland, Miocene lignites and kaolinite outcroppings were not completely removed by the overriding ice. The Winooski River, draining 2900 km^2 of northern Vermont, cuts a narrow valley more than 1000 m deep, directly across the grain of the Green Mountains. This drainage, and the large-scale topography we see today, almost certainly existed prior to glaciation.

Drainage in most of northwestern Vermont is generally toward the Champlain Lowland, and from there to the St. Lawrence River. When ice filled all or part of the Champlain Lowland (both during glacial advance and retreat), north-flowing drainage was blocked, glacial lakes were impounded in the lowland, and water flowed south to the Hudson or east to the Connecticut River (Fig. 2). Surface water would have been impounded in all the major drainages, forming lakes and backing up into tributary valleys.

As ice melted and the ice margin retreated, what began as isolated ice-marginal lakes coalesced into glacial Lake Vermont, which drained into the Hudson River Valley through a spillway near the southern end of Lake Champlain (Chapman, 1937). Lake Vermont at its lowest stage held approximately 240 km^3 of water, about ten times the volume of present-day Lake Champlain (Desilets and Cassidy, 1993; Fig. 2B). Lake Vermont ended when ice-margin retreat allowed marine waters of the Champlain Sea into the isostatically depressed Champlain Lowland (Fig. 2C). Over the next 1000 to 1500 years, isostatic rebound increased the elevation of the Richelieu River sill at a rate greater than eustatic sea-level rise, finally isolating Lake Champlain from marine waters and freshening the lake by about 10 ky - ^{14}C BP (Parent and Occhietti, 1988).

The Winooski River and its tributaries were directly influenced by falling base levels during the late Pleistocene and early Holocene (Fig. 3). Valley fills, formed during early, higher base levels, were rapidly incised and left as terrace remnants when glacially-impounded lakes drained and base levels not only fell, but were located increasingly further from tributary valleys (Whalen, 1998). Alluvial fans formed on some of the abandoned terraces, preserving evidence of past sedimentation events. In some locations, these fans archive up to an 8000 ^{14}C yr record of hillslope activity (Bierman et al., 1997).

Human activity has significantly affected the Vermont landscape. Vermont was first settled in the late 1700s. By 1850, much of Vermont had been cleared for agriculture. Initial clearance was for cropland; later clearance was for sheep grazing. In the 1870s, land at higher elevations was cleared for timber. Landscape response to this clearance is well preserved in the geologic record. Hillslopes became unstable and sediment yield appears to have increased significantly as documented by aggrading alluvial fans and flood plains (Bierman et al., 1997 and Fig. 4).

This trip provides field examples illustrating what we do and do not know about the northern Vermont landscape. This guidebook provides general background information that will be supplemented by daily handouts. These handouts will provide logs of the many trenches that will not be opened until soon before the field trip.

This guidebook draws heavily on the past five years of work by students and faculty at the University of Vermont and elsewhere. Undergraduate thesis students Paul Zehfuss (UVM BS 1996) and Kristine Bryan (UVM BS 1995) mapped alluvial fans and glacial deposits, respectively. Tim Whalen (UVM Geology MS 1998) and Chris Valin (UVM Geology BS 1997) surveyed and trenched along three main tributaries of the Winooski River. Lin Li (UVM Geology MS 1996) first cored Ritterbush Pond and conducted palynological analysis of her cores. Sarah Brown (currently completing UVM Geology MS) studied the sedimentology of three additional Ritterbush Pond cores in great detail. Amy Church (UVM Geology MS 1997) first trenched the Moultroup fan. Students from the UVM Field Naturalist program, Mike Loso and Lyn Baldwin, surveyed topography and trees. Undergraduates in UVM Geomorphology and Geohydrology classes, named in the text where appropriate, found and first interpreted many of the sites referred to in this guidebook. Tom Davis (Bentley College) provided field and laboratory experience with pond sediment cores and John Southon (Livermore National Laboratory) is responsible for AMS dating.

DAY ONE

Today we will visit outcroppings and trenches of glacial marine, glacial lacustrine, fluvial, and alluvial fan sediments. We will see evidence for the channeling of ice in river valleys during deglaciation and the resulting distribution of glacial, immediately post glacial, and Holocene sediments. We will examine trenches cut into several river terraces and alluvial fans.

Stop 1-1 — Town Line Brook

Town Line Brook is a small, deeply incised tributary of the Winooski River (Fig. 5). Its valley walls expose up to several meters of fluvial gravel cut into tan, well-sorted fine to very fine sand that overlies gray silt with interbedded fine sand. Groundwater usually seeps from the contact. Underlying these materials are thin silt/clay couplets (rhythmites) near the base of the exposure. The rhythmites were deposited in the quiet waters of Lake Vermont, whereas the gray silt and sand were deposited in the Champlain Sea (Figure 2). The Champlain Sea sediments can be distinguished by the absence of rhythmites and the presence of small, white bivalves (*Macoma Baltica*), which can occasionally be found on this outcrop. The sands signal the encroachment of the paleo-Winooski River delta to the area in Champlain Sea time. The gravels fill channels in the fine sand and are distributary channels of the paleo-Winooski River delta.

The valley of Town Line Brook has been widened by landsliding and deepened by fluvial incision — controlled to a great extent by the distribution of ground water. The fine-grained sediments contain sandy interbeds along which ground water preferentially flows. The interbeds wash out, causing small-scale slumps and toppling failures of the more cohesive,

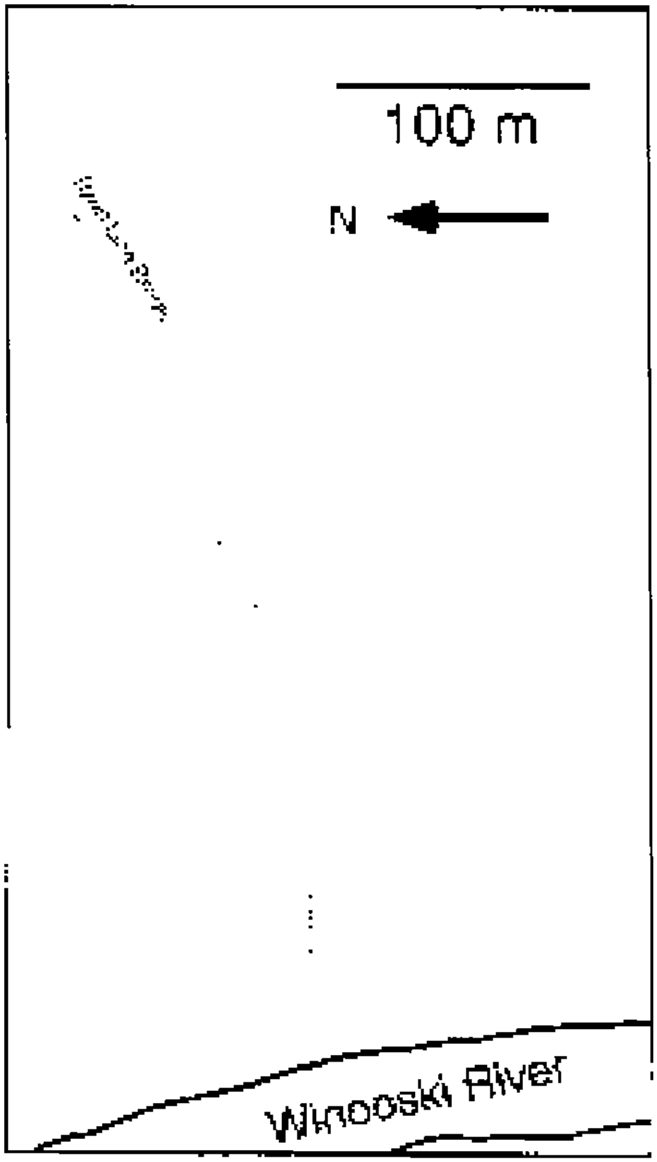


Figure 5: Topographic map of Town Line Brook area. Scale bar is 200 m. Adapted from U.S.G.S. Burlington quadrangle, 1:24,000, original map 1948, photorevised 1987.

overhanging fine-grained material, which liquefies easily (try stamping on some failed material). The liquefaction is important because it allows failed material to be evacuated easily from the valley by rather modest stream flows. Once the fine-grained deposits have failed, the overlying, non-cohesive, and permeable deltaic sand and gravels also fail by translation and toppling. Such failures are particularly common during wet periods when the water table rises further into the sand.

Today, there are only a few active slides, but slide scars are prevalent along the watercourse. Ring count-

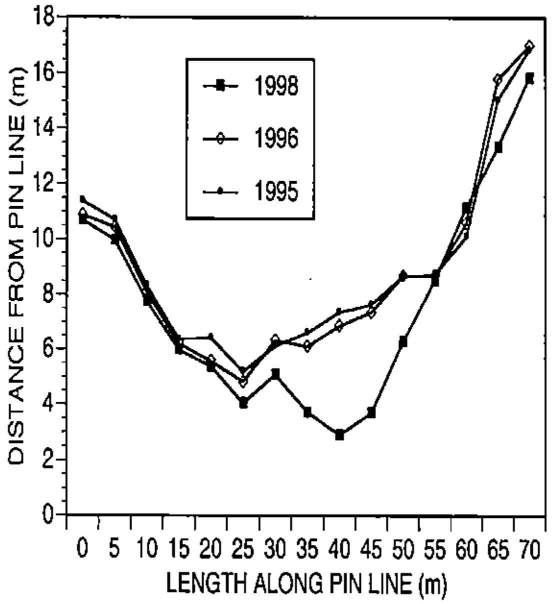


Figure 6: Pin line data for the main Town Line Brook landslide showing retreat of landslide scarp over the past three years. Data gathered by successive UVM Geohydrology classes from 1995 until 1998.

ing of tree cores shows that most of the trees within the currently inactive slides are < 100 years old (Baldwin et al., 1995). The age of these trees suggests that Town Line Brook hillslopes began to stabilize in the late 1800s, coincident with the reforestation of northwestern Vermont (Fig. 4). According to local residents, the major landslide complex became active within the past 20 years. Pin line measurements suggest that the scarp has been retreating episodically over the past three years at rates of several cm to >1 m yr^{1} (Fig. 6). Using the geometry of the slide, one can estimate that this slide alone provides 150 to 250 m^{3} yr^{1} of sediment to the Winooski River. A long-term average rate of sediment export from Town Line Brook (10 to 15 m^{3} yr^{1}) can be calculated using valley volume (about 100,000 to 150,000 m^{3}) and assuming that the paleo Winooski River delta was abandoned 10,000 years ago when the Champlain Sea drained.

Stop 1-2 — Jonesville Rock and Ice Retreat

This outcropping of Underhill formation mica schist preserves striations and groves indicating that the last ice flowing over this outcrop moved from N 70° W. This flow direction is parallel to the orientation of the Winooski River Valley and indicates that ice flow was channeled, presumably during retreat, by this major topographic feature. Measurements of striae as a func-

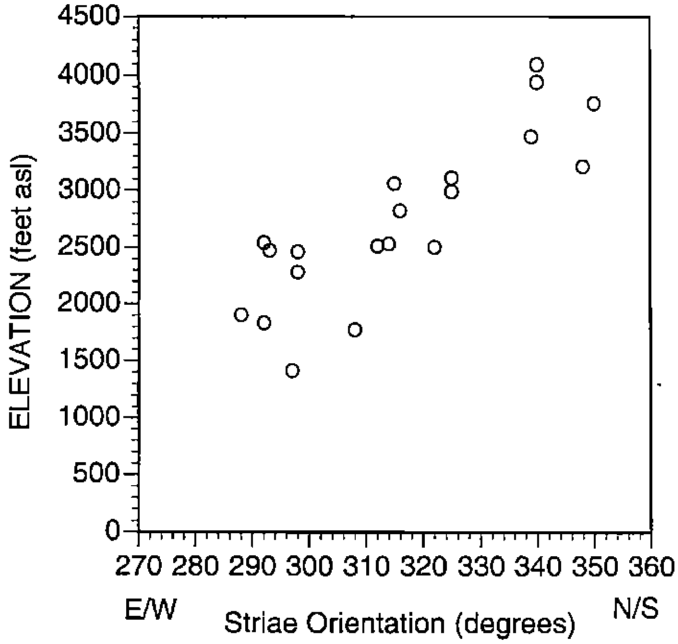


Figure 7: Orientation of striae as a function of elevation, transect from near Jonesville Rock to the summit of Camels Hump; figure from Malchyk and Kelly (1996).

tion of elevation above the valley bottom (Malchyk and Kelly, 1996) show that striae become oriented toward regional flow directions (approximately NNW-SSE) at higher elevations on Camel's Hump (Fig. 7).

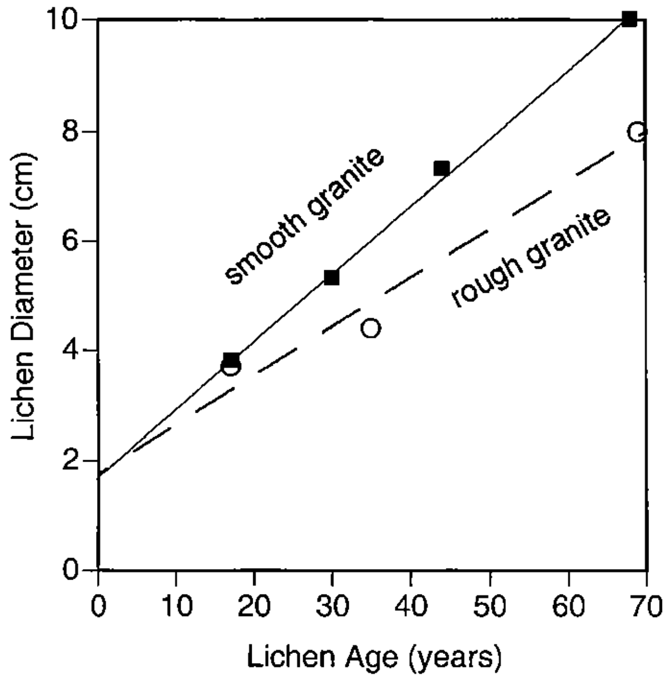


Figure 8: Calibration of lichen (probably *Xanthoparmelia plittii*) maximum diameter for the Richmond, Vermont area (Sundae, 1997).

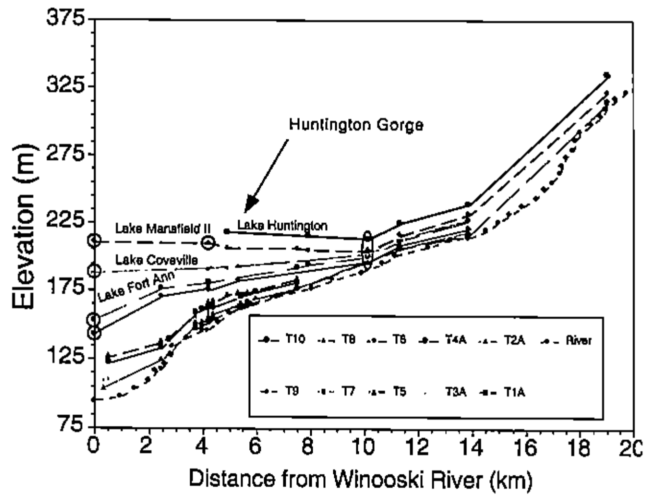


Figure 9: Longitudinal profile of terraces bordering the Huntington River Valley. Adapted from Whalen (1998, Figure 4.2). Figure is vertically exaggerated.

Sundue (1997) measured lichen size as a function of underlying tombstone age in the Richmond, Vermont cemetery. He established that lichen growth rates over the past century are linear and on the order of 1 mm yr^1. This rate is similar to that determined for lichens on tombstones less than 50 years old in the Champlain Lowland (Royce and Young, 1994). Using the Richmond calibration (Fig. 8), lichen diameters on bare, striated bedrock outcrops similar to and near the Jonesville Rock, suggest exposure within the last century or two. Such recent exposure is consistent with the excellent preservation of striae on this relatively easily weathered rock. Exposure of the bare rock surface was most likely the result of land clearance for farming and grazing during the 1800's.

Stop 1-3 — Huntington Gorge

Huntington Gorge is cut through schist of the Underhill Formation and displays well-developed pot holes and plunge pools. The gorge appears to exploit joint sets trending N 82° E, N 75° W, and N 51° W (Christman and Secor, 1961). Until recently, it was not known when the gorge formed although the common speculations include incision immediately after deglaciation when poorly vegetated slopes generated large amounts of sediment-charged runoff or catastrophic draining of a glacial lake. However, recent work by Whalen (1998), who surveyed longitudinal profiles of the Huntington River terraces, 14C-dated terrace sediments, and correlated terraces to changing base-levels, constrains the age of the present Huntington Gorge.

Whalen's T6 terrace passes over the gorge with no apparent increase in gradient indicating that during T6 time, the gorge was not exposed (Fig. 9). The T6 terrace was graded to the Fort Ann stage of Lake Vermont, which ended 11.7 14C y BP with the initiation of the Champlain Sea. Thus, 11.7 14C y BP, is the upper limit for the age of the gorge. The gorge may have been first exposed as late as 8500 14C y BP, the oldest age for charcoal pulled from the overbank sediments of terrace T5, the first terrace showing a gradient increase in the area of the gorge (Fig. 9). These dates lead to an important conclusion; the gorge was formed by fluvial erosion at least 1000 14C years after local deglaciation, not by the catastrophic draining of any lake.

Stop 1-4 — Huntington River Terraces

The Huntington River is bordered by 10 levels of terraces representing a steady fall in base level and the incision of glacial and post-glacial valley fill. The terraces, their elevation, spatial distribution, and stratigraphy have been surveyed and are discussed in detail by Whalen (1998).

Background

Terraces in the Huntington River Valley range in age from >12,500 ^{14}C years to historic. The highest and oldest terraces represent early post-glacial stream deposits graded to the level of an isolated proglacial lake impounded in the southern Huntington River Valley by ice to the north (Fig. 10, Stage I). Water from this isolated lake poured south through an unnamed 460 m (1510 ft) spillway at the southern end of the valley. As the ice continued to melt and its margin retreated northward, a lower spillway through Hollow Brook was uncovered (204 m, 670 ft) and water in the Huntington River Valley flowed west into Lake Vermont. Thus, the level of the Huntington River Valley lake dropped 255 m and the lake shrank dramatically (Fig. 10, Stage II).

Further ice retreat radically changed the drainage patterns of northern Vermont. As soon as the ice margin retreated enough to expose the col at Gillett Pond (Figure 10, Stage III), flow directions in the Winooski River Valley reversed! The level of Lake Mansfield I, which filled the Winooski River Valley, dropped 50 m from 279 m (controlled by the Williamstown spillway)

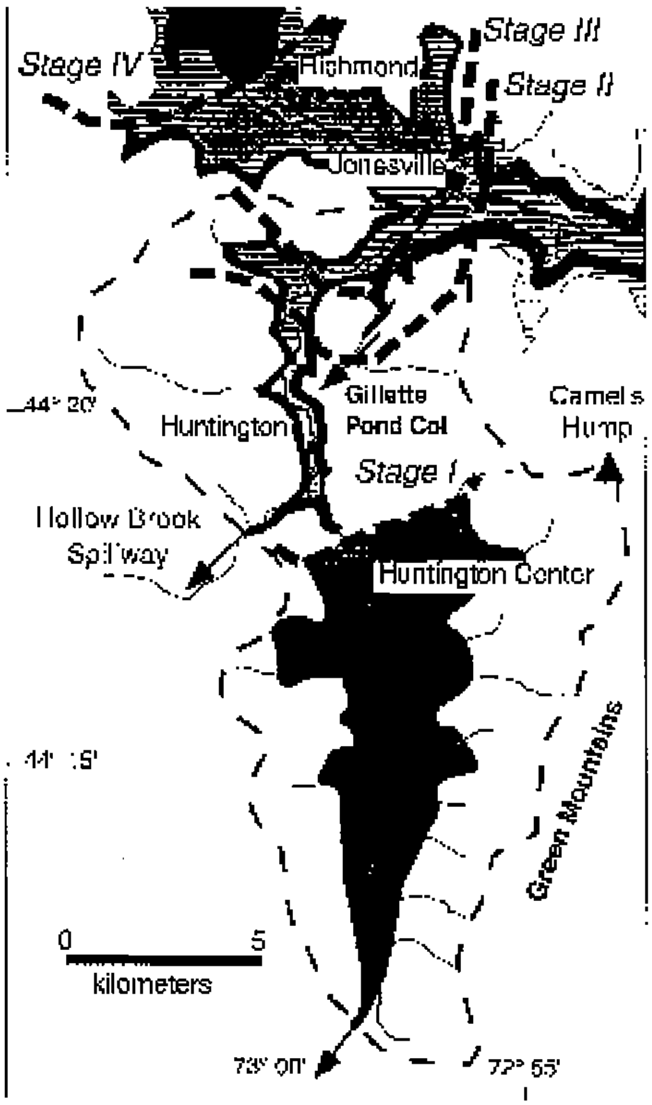


Figure 10: Map of ice margins and proglacial lake extents in the Huntington River Valley of northwestern Vermont. Hollow Brook, Gillett Pond, and southern Huntington River Valley thresholds are shown with arrows indicating flow directions. Ice margins are shown schematically and labeled by stage. Stage I is the earliest stage. Lake levels impounded by ice are shown in varying shades of gray. From Figure 6 of Wright et al. (1997).

to 229 m (controlled by the Gillett Pond spillway, see Figure 11 and Larsen, 1987b). Waters of the Winooski River Valley's Lake Mansfield, which had previously poured into Lake Hitchcock and drained through the Connecticut River Valley, now poured into Lake Vermont and from there to the Hudson River. The once-isolated, local lake in the Huntington River Valley, with a drainage basin of perhaps 100 km^2, now received water and sediment from the Winooski River drainage basin, more than 20 times larger.

As ice retreat continued, the Gillett Pond spillway was abandoned and lakes in Huntington and Winooski Valleys merged into Lake Mansfield II (Fig. 10, Stage IV and Fig.11). Water and sediment from this system continued to pour over the Hollow Brook spillway into the Champlain Basin, building the massive delta into Lake Vermont that we will see at the end of the day (stop 1-6). Water from the Winooski Valley continued to flow south through the Huntington River Valley until the ice margin had melted back far enough into the lowland that the lake in the Winooski River Valley (Lake Mansfield II) and the lake in the Champlain Valley (Lake Vermont) merged (Fig. 2B and Fig. 11, Stage V). At this time, the Hollow Brook spillway was abandoned, lake waters withdrew from most of the Huntington River Valley, and water in the Huntington River Valley flowed north. The current drainage system had been established.

Fluvial History

The fluvial history of the Huntington River Valley begins at different times in different places. In the upstream reaches of the valley, the fluvial history began as soon as the Hollow Brook spillway opened and the high-level lake (Fig. 10, stage I) drained. In the lower reaches of the Huntington River Valley, the fluvial history could not begin until Lake Vermont lowered at the beginning of Champlain Sea time.

A general history of valley evolution is illustrated in Figure 3. Initially, the river flowed over and augmented glacial-lacustrine fill terraces (Fig. 9, T-10, T-9 and T-8). These deposits are generally a coarsening-upward sequence of rhythmically bedded silt grading into sand (some of which may locally be deltaic in origin and therefore exhibit foresets). The fill terraces are capped by fluvial deposits. Although these fill terraces have been mapped as deltas based on their morphology (i.e., Wagner 1972), many are located far from incoming streams.

The highest extensive terrace, T-9, was built during Stage IV when lakes in the Huntington and Winooski River Valleys had coalesced. At this time, water in the Huntington Valley generally flowed north to south from the Winooski River Valley to the Hollow Brook spillway. In the portion of the valley once occupied by the proglacial lake, terrace T-9 now slopes to the south between 0.25 and 0.67 m km^1, depending on the orientation of the northward-sloping Huntington River Valley. The T-9 terrace slope reflects either pure isostafigure 11

Figure 11: Schematic cross-sections through the Huntington River Valley as ice retreats and lake levels lower. Adapted from Waalen et al. (1998, Figure 7)

Table 1: Summary of Terrace Characteristics at Moultroup and Aldrich Farms

Terrace	Type	River System	Elevation	Incision (m)	Age ^{14}C y BP (m)
T8	Fill Terrace	Meandering	190.98	-	12.8-12.6
T7	Fill Cut Terrace	Braided	180.44	10.54	12.6-11-7
T6	Fill Cut Terrace	Braided	176.25	4.19	11.7-10.8
T5	Fill Cut Terrace	Meandering	165.57	10.68	10.8-7.8
T4	Fill Cut Terrace	Meandering	162.50	3.07	7.8-60 (?)
T3	Fill Cut Terrace	Meandering	158.90	3.60	6.0(?)-2.5
T2	Fill Cut Terrace	Meandering	155.47	3.43	2.5-02
T1	Fill Cut Terrace	Meandering	151.25	4.22	0.2 Today

Note: Ages based on correlations to dayed base levels in Champlain Valley and dated alluvial fan and terrace deposits (Table 2; Whalen, 1998).

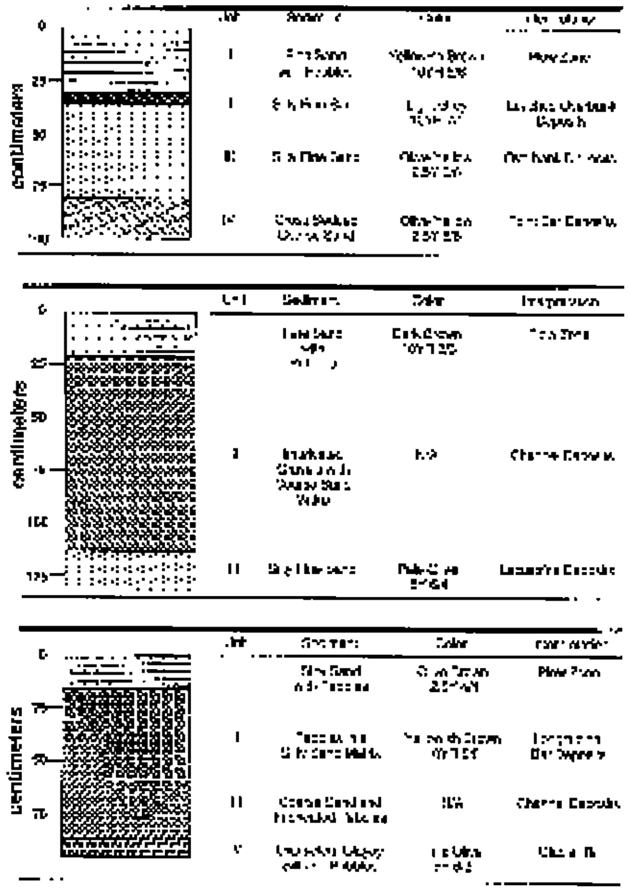


Figure 12: Examples of sedimentological difference in Huntington River terrace deposits. From Whalen et al. (1998).

tic rebound of an initially flat surface or isostatic tilting enhancing the gradient of a surface that originally sloped southward toward the spillway at Hollow Brook.

The lower reaches of the next terrace, T-8, were also graded into a proglacial lake, Lake Vermont (Coveville Stage). This terrace slopes northward today, suggesting that it was deposited by waters flowing north down the Huntington River Valley toward Lake Vermont. With the retreat of ice, lake levels had lowered and the Huntington River had re-established its northerly course. Later terraces are inset within the higher terraces and each slopes downstream more steeply than the terrace above; these data imply that the gradient of the Huntington River has increased since deglaciation.

Incision and gradient changes were a response to base level drops in the lower reaches of the Winooski Basin. The amount of incision varied along the Huntington River Valley and depended on two factors: the magnitude of base-level change and the distance from the base-level change. Base-level changes began with the lowering of proglacial lakes and continued with the decline of the Champlain Sea.

Each drop in water level initiated a wave of incision that propagated upstream until a bedrock knick point was encountered or until the wave of incision diffused due to a lack of stream power. Incision began by cutting of the valley fill, followed by a period of relative stability and flood plain development; therefore, cut-fill terraces were formed. In the Huntington River Valley, regional base-level changes ceased to be the direct cause of incision after Champlain Sea time because bedrock knick points were exposed on the lower reaches of the Winooski River.

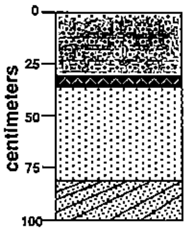
Subsequent to the incision driven directly by base-level change, episodic incision continued to occur, perhaps mediated by climate change and/or internal adjustments of the river system. Terraces formed later in the Holocene have less relief between levels and different morphology. Younger terraces have surfaces dominated by ridges and swales.

Moultroup and Aldrich Terrace Sequences

The flight of terraces at the Moultroup and Aldrich farms provides an excellent example of the differences between terraces formed by base level-driven incision and terraces formed by internal adjustments. Eight terraces are identified here representing three stages of valley evolution, evident by both the relief between terraces and the stratigraphy of terrace deposits. The first stage of the valley evolution is the deposition of the glacial-lacustrine fill terrace (T8). The second stage is dramatic incision followed by deposition of cut-fill terraces as a response to base level changes (T7-T6). The final stage is periodic incision related to internal adjustments of the fluvial system, perhaps triggered by environmental changes (T5-T1).

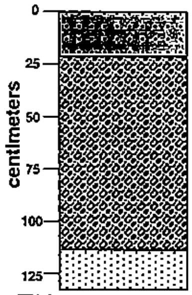
Terrace ages are constrained in different ways: directly from organic matter that they contain, indirectly by correlation to base levels dated by other means, and by the limits imposed from dated, overlying alluvial fans (Table 1 and Whalen, 1998). We have direct dates for a number of the lower terraces (Whalen, 1998). Two pieces of charcoal from the terrace deposits of T6 yielded ages of between 8100 and 8300 ^{14}C y BP and provide a minimum age for the formation of T6. However, the age of T6 is believed to be much older based on: 1) correlation to the Upper Marine stage of the Champlain Sea, 2) the ages of the fans built on T5, and 3) the age of the T5 deposits.

T8



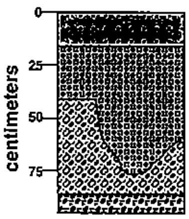
Unit	Sediment	Color	Interpretation
I	Fine Sand with Pebbles	Yellowish Brown 10YR 5/6	Plow Zone
II	Silty Fine Sand	Light Grey 10YR 7/1	Leached Overbank Deposits
III	Silty Fine Sand	Olive Yellow 2.5Y 6/6	Overbank Deposits
IV	Cross Bedded Coarse Sand	Olive Yellow 2.5Y 6/6	Point Bar Deposits

T7



Unit	Sediment	Color	Interpretation
I	Fine Sand with Pebbles	Dark Brown 10YR 3/3	Plow Zone
II	Imbricated Gravels with Coarse Sand Matrix	N/A	Channel Deposits
III	Silty Fine Sand	Pale Olive 5Y 6/4	Lacustrine Deposits

T6



Unit	Sediment	Color	Interpretation
I	Silty Sand with Pebbles	Olive Brown 2.5Y 4/4	Plow Zone
II	Pebbles in a Silty Sand Matrix	Yellowish Brown 10YR 5/8	Longitudinal Bar Deposits
III	Coarse Sand and Imbricated Pebbles	N/A	Channel Deposits
IV	Diamicton: Clayey Silt and Pebbles	Pale Olive 5Y 6/3	Glacial Till

Fig 12

A basal date obtained from wood below the Audubon fan on T5 is 8530 ± 100 ^{14}C y BP (CAMS# 20901) and another basal age from wood found 4.0 m below the Moultroup fan surface across the valley on T5 is 7835 ± 105 ^{14}C y BP (GX-20276). These ages agree well with the age of 7790 ± 60 ^{14}C y BP (CAMS# 30353) from a piece of charcoal recovered from 0.75 m below the T5 surface, suggesting that T5 was formed by 8500 ^{14}C y BP and was active until at least 7800 ^{14}C y BP.

The age of 2500 ± 60 14C yBP (CAMS# 22994) from 4.0 m below Aldrich fan C surface on terrace T2A places an upper limiting age on its formation. The younger alluvial fan, Aldrich B, with a basal age of 1900 ± 50 ^{14}C y BP (CAMS#30358), is located just downstream on terrace T2B. The difference in ages between these adjacent fans points either to the migration of the river channel during this period or a difference in the timing of fan initiation. A piece of cloth, an historical artifact, was recovered from the T1 deposits. The Aldrich A fan built on T1 is <100 ^{14}C y BP (GX-2139), consistent with an historic age for T1.

Cross-valley profiles of the terraces clearly show that the magnitude of incision, represented by the relief between terraces, is larger for the highest terraces and decreases towards the river . The magnitude of incision between both T8/T7 and T6/T5 is over 10 m, and for all terraces below T5, less than 5 m of incision took place (Table 1). The obvious differences in relief between the upper (T8-T5) and the lower (T5-T1) terraces corresponds to the change in incision mechanism and river behavior that is also evident in the terrace stratigraphy.

Backhoe trenches and shovel pits, opened on nearly every terrace here, revealed three different terrace stratigraphies (Fig. 12). Common to all terraces is the presence of a 25-cm-deep plow zone capping the fluvial deposits. The plow zone represents the historical land use at this location (pers. comm., Henry Moultroup). The upper meter of the glacial-lacustrine fill terrace, T8, is composed of cross-bedded sand overlain by silty fine sand, which suggests deposition by a meandering fluvial system. The stratigraphy of the pits and trenches from T7 and T6 (represented by the T6 log in Fig. 12) is consistent with a braided fluvial system. Two fluvial units, channel deposits of imbricated pebbles and longitudinal bar deposits of pebbles and sand, overlie a diamicton interpreted to be glacial till, but overbank deposits are thin or absent on T7 and T6. The terrace deposits from T5 to T1 all have a basal unit of imbricated gravels, deposited in the channel, overlain by laterally accreted sand that is capped by overbank deposits of silty fine sand. This stratigraphy implies that the fluvial system behaved similarly to today's meandering river since the time of T5.

The terraces at the Moultroup and Aldrich farms provide one of the best examples of the different forms of fluvial terraces found in the Winooski basin. The terrace chronologies discussed here are consistent with those studied in the Little and Mad River Valleys (to the Northeast and East, respectively) and may be applicable to other terraced valleys in Vermont (Whalen, 1998).

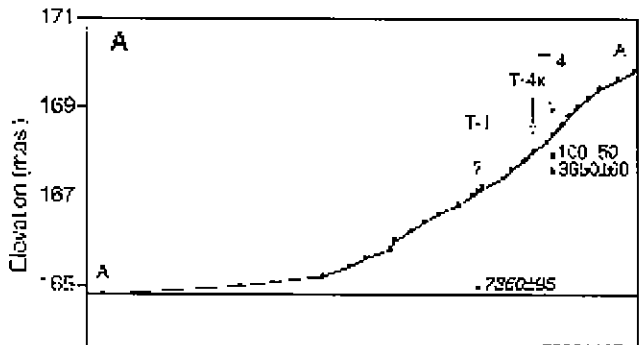


Figure 13: Moultroup (MUL) alluvial fan . A. Vertically exaggerated cross-section of fan showing location of radiocarbon-dated wood (italic) and charcoal (not italic). Open circles indicate sample locations. B. Estimated aggradation rates in ^{14}C years (assuming fan was right, circular cone) demonstrate response of fan during historic period (180 years, based on settlement history).

Stop 1-5 — Aldrich and Moultroup Alluvial Fans

Alluvial fans are well developed on and below some terraces of the Huntington River. These fans were first studied by Zehfuss (1996) and Church (1997) who opened 11 trenches on five fans. All of the fans are small (<2500 m^2). They contain 2000 to 4000 m^3 of sediment and range in age from >8000 ^{14}C years to historic.

Introduction

Alluvial fans are seldom-studied landforms in the Northeast. We have investigated 22 alluvial fans in the Huntington River Basin (Bierman et al., 1997; Church and Bierman, 1994, 1995; Zehfuss and Bierman, 1995; Zehfuss, 1996), clustered in the northern part of the river valley where Huntington River terraces provide a platform for fan sedimentation. We have studied intensively three sites (Audubon, Moultroup, and Aldrich) and have used these sites to infer the regional history of hillslope stability during Holocene.

Most fan sediments are poorly sorted, although there are occasional thin (<10 cm) beds of well-sorted, clast-supported gravel, as well as black laminae that may represent decomposed leaf mats or concentrations of finely disseminated charcoal. Material deposited in the fans originates from distinct drainage basins located either on the surrounding mountain sides, or on slopes created by the risers of higher Huntington River terraces.

We have radiocarbon dated 14 samples of wood and charcoal from five alluvial fans in order to determine the timing of aggradation. These data show that two fans began to aggrade in the early Holocene (between 8530 and 8060 ^{14}C y BP, and between 7835 and 7360 ^{14}C y BP), two in the late Holocene (2500 and 1900 14C y BP), and one fan aggraded over 4 m during historic time (<100 ^{14}C y BP). During the field trip, we will view the fans from which we have obtained aggradation histories.

Each fan contains a well-preserved, buried soil profile, generally within the uppermost meter, but in one fan, the soil is buried by four meters of debris. Radiocarbon ages confirm that the sediment above the buried soils was deposited coincident with historic destabilization of hillslopes. Our data suggest that the extensive deforestation that occurred during settlement in the late eighteenth century destabilized the hillslopes and increased fan aggradation rates, burying prehistoric soil profiles. Without stability provided by root systems, the soils and sediments were more easily eroded from drainage basins, increasing sedimentation rates on the fan surfaces. After initial deforestation, sediment continued to be deposited on the fans at a rapid rate. The recovery of forest growth on the hillsides above Vermont fans has returned some, but not complete, stability to slopes.

A plow zone is usually present above the buried soil, representing the time following deforestation during which the fields were used for farming. The abundance of organic material in the fan deposits, such as wood and charcoal, facilitates the use of radiocarbon dating for obtaining ages of sediment and estimating rates of deposition.

Moultroup Fan

The Moultroup fan is on the west side of the Richmond–Huntington Road, just south of the right-hand turnoff onto Dugway Road. There is a second, larger fan to the north, with a house built on its surface. The Moultroup fan was trenched extensively in the summer of 1994 and the fall of 1997. The fan radius is approximately 65 m, with an apex height of 5 m (Fig. 13A). A vegetation change to tall grass and reeds in the center of the field marks the approximate toe of the fan, a seasonally-saturated area of groundwater discharge. The Moultroup fan has accumulated sediment derived from glacial till and glaciolacustrine deposits that cover the adjacent bedrock highlands.

Trenching of the Moultroup fan reveals a complex stratigraphy. The slightly asymmetrical appearance and surface morphology of the fan suggest that it may be composed of multiple lobes. On the Moultroup fan, there is sufficient radiocarbon and stratigraphic data to estimate sediment accumulation rates over much of the Holocene (Fig. 13B). Aggradation rates were rapid in the early Holocene when the fan was beginning to form, between 7835 to 7360 ^{14}C y BP. Sedimentation slowed considerably in the mid-Holocene (7360 to 100 ^{14}C y BP). The stratigraphy of the Moultroup trenches reveals a distinct soil profile near the fan surface buried by about 0.5 to 0.9 m of poorly sorted sediment. A charcoal age (<100 ^{14}C y BP) just above the buried soil and a horseshoe found within the soil indicate that the overlying sediment post-dates European settlement.

The Aldrich Fans

The Aldrich fans are found on the south side of the Huntington River along Dugway Road, about 1 km east of the Moultroup Fan. From the river looking south, two fans are readily apparent in the center of the far side of the Aldrich field. The fans extend almost 80 meters into the field, and coalesce in the mid-fan portions. Farther east, the smaller Fan A is less obvious.

A steep slope, formed by a higher Huntington River terrace (T-8), is the source of fan sediment. Drainage basins cut into this higher terrace contain ephemeral streams leading to the fans below and are filled with ice and/or flowing water in the winter and spring months. Small stream beds on the terrace top converge at the head of the drainage basins. Clasts ranging in size from large cobbles to small pebbles line the ephemeral stream beds leading to the fans. Deposits of sand and

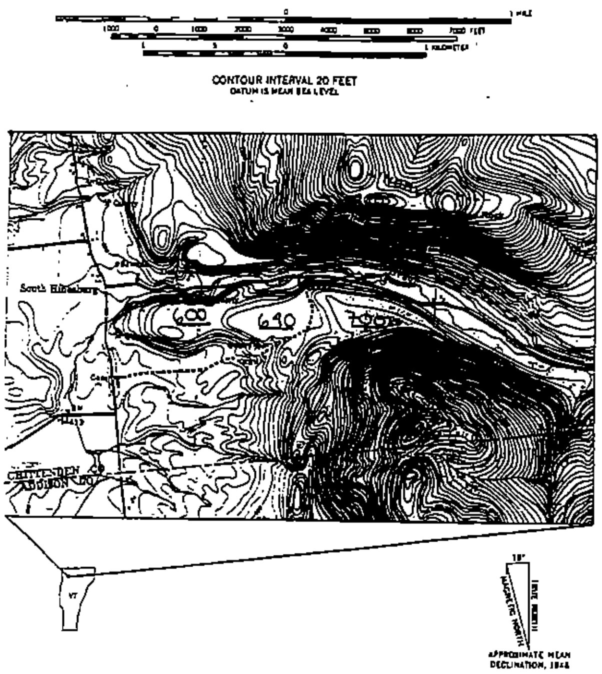


Figure 15: Topographic map of the Hollow Brook delta-complex showing levels graded to 213 m (700 feet), 195 m (640 feet), and 183 m (600 ft) elevation. Field trip stop will likely be in a public gravel pit to the north of Hollow Brook. The exposure we will examine is cut into the delta built to the 183 m level. From USGS 1:24,000 Hinesburg quadrangle, 1948.

silt are exposed along the steep sides of the drainage basins, and in some places have plunged into the stream beds as small landslides (5–10 m^3) containing wet and cohesive sediment. The scarps of the slides appear as hollows below a thick (≈ 25 cm) mat of vegetation.

Soil trenches and communication with residents of the Huntington and Richmond regions verify that the Aldrich field was plowed repeatedly prior to 1960. A cross-section of Fan B (Fig. 14) shows the location of the plow zone in the trenches. Towards the toe of Fan B, the paleosol A horizon is incorporated into a cumulative plow zone. It can be inferred that the paleosol A horizon existed throughout the entire Aldrich field at one time, but was completely mixed by plowing, where it was not buried deeply enough to be preserved by alluvial fan sediments. Following the cessation of row-crop farming on the Aldrich meadow and its subsequent and continued use for unimproved pasture, the cumulative plow layer has been shallowly buried in places by intermittent fan deposition.

On the Aldrich fans, thin (<1 mm) beds of fine silt (never thick enough to cover the grass) have been deposited on small parts of the fan surface. These deposits resulted from ephemeral stream activity during severe storms and winter thaws within the past few years. On these same fans, record rainfalls during July, 1998 triggered deposition of gravel on the fan apex and sand and silt further downfan.

Stop 1-6 — Hinesburg Delta, Hollow Brook Spillway, and Lake Vermont Levels

Hollow Brook is a dramatically underfit stream. It occupies a wide (up to 450 m) and deep (up to 210 m) valley sculpted into bedrock. The valley floor (204 m, 670 ft asl) is flat and poorly drained. Once ice retreated north and west of Hollow Brook, the valley served as a spillway for waters impounded in the Huntington River Valley. Because the floor of Hollow Brook Valley is only about 7 m above the current level of the Huntington River, it is possible that at times in the past (last interglacial?), the Huntington River entered the Champlain Lowland directly through the Hollow Brook Valley.

At the mouth of Hollow Brook, where it dispatches into the Champlain Lowland, is a large and thick accumulation of sorted sediment. Access to these sediments is limited because the owner of the largest gravel pit refuses entrance to geologists based on liability concerns. Extensive gravel mining over the past several decades has removed or altered much of the original topography, but older aerial photographs and topographic maps (Fig. 15) show a series of inset surfaces at elevations of 213 m (700 feet), 195 m (640 feet), and 183 m (600 ft). The lower two surfaces (195 m and 183 m) are graded to the Coveville and a post-Coveville Stage of declining Lake Vermont. The upper surface (213 m) must correspond to a local ice-marginal lake because the 213 m level does not correspond to any extensive lake stage in the Champlain Lowland and is 10 m above the current elevation of the Hollow Brook spillway.

Kjelleren (1984) and Bryan (1995) suggest that much of the deposit is deltaic. This conclusion is based on the observation of westerly flow directions in sediments and the predominately eastern provenance of clasts contained in the deposit. Earlier mapping by Stewart and MacClintock (1969) had suggested that the delta was a kame terrace. Our observations of flow directions, made over the past five years at the 183 m level, are consistent with a deltaic origin for at least the lower part of the landform. Sediments that make up the highest surface (213 m) are poorly exposed. This surface likely represents a deposit built into a small, isolated lake, marginal to ice in the Champlain Lowland. It is not clear whether the sediment in this high-level deposit originated from Champlain Valley ice, or ice in the Huntington River Valley.

FRIDAY EVENING, DAY 1 — PERKINS GEOLOGY BUILDING

During and after dinner on Friday, there will be several activities occurring in the Geology Building including core opening, lab tours, web page review, and email access.

Ritterbush Pond Cores

Ritterbush Pond is a small (300 m wide, 14 m deep) body of water set into a rock basin at 317 m asl. It was first cored by Sperling et al. (1989) who were attempting to date what they considered to be the first sedimentation in a tarn formed by receding, post-Laurentide alpine glaciers. Their core was taken near the lake margin and gave a lowermost age of 10,090 ^{14}C y BP on bulk organic material. Ritterbush Pond was re-cored in its deepest part by Li (1996), whose basal age of 11,940 ^{14}C y BP is considerably older. Li counted pollen in two overlapping Livingston cores from the pond and first identified discrete layers of organic-poor sand, silt, and clay.

Over the past two years (S. Brown, UVM MS candidate) has studied, in detail, three additional cores from Ritterbush Pond (Brown et al, 1997, 1998). These cores record a history of Holocene hydrologic change, presumably a response to changing climate. A total of 28 ^{14}C ages document the timing of several dozen silt and sand layers in the otherwise organic -rich cores. The sand and silt deposits cluster temporally (2500, 6000, and 8500 ^{14}C y BP), suggesting fluctuations in sediment dynamics since stabilization of the 2.2 km^2 watershed, around 9000 ^{14}C y BP. The organic material in the sand and silt layers has carbon and nitrogen stable isotope values supportive of a terrestrial origin, but charcoal data do not support slope clearing fires as a trigger for erosion. The clustering of event ages suggests that erosion and sedimentation were likely triggered by changes in storm frequency and/or intensity.

"Vermont Landforms" Web Page

In the attempt to further the connection between university geologic research and public outreach, we have developed an interactive educational web site for high school/junior high school students. The page, *Vermont Landforms* uses recent University of Vermont research in New England geomorphology as the basis for a set of learning tools. The goals of the page are: 1) to introduce the process of scientific inquiry, 2) to introduce geomorphic principles and Vermont geology, and 3) to integrate the latest geologic research into secondary education level classrooms.

This site has four main elements: 1) the main learning modules, 2) the supporting learning modules, 3) the historic photo database, and 4) the glossary. The main learning modules are based on current graduate and faculty research and are based on the Kolb model modified for classrooms activities (Svinicki and Dixon, 1987). These modules are composed of cycles: short information sections, virtual experimentation, virtual measurements, problem sets, and discussion sections. The supporting learning modules consist of background information intended to support the main modules. These support modules are accessible through icons found everywhere in the site. There are currently three support modules: 1) Vermont Bedrock Geology, 2) Vermont Glacial Geology, and 3) Post-glacial Geology in Vermont. At any point, the page allows one to review previously covered material or to continue on to the next section.

The third segment of this web site is the historic photo database that is continually being updated. This is a database containing historic landscape photographs of various places in Vermont accessible through a map-index of counties (and towns). This section is also accessible from anywhere in the site. The last element of the site is a graphic/text glossary accessible by clicking on highlighted terms within the text of the site. Eventually, there will be a "FOR TEACHERS ONLY" section containing curriculum materials to help teachers use the site more effectively within their classrooms.

DAY TWO

Today we will cross the Green Mountains and climb Mt. Mansfield, the highest peak in Vermont. We will examine both Holocene and late glacial sediments, walk on an active alluvial fan, and conclude the day

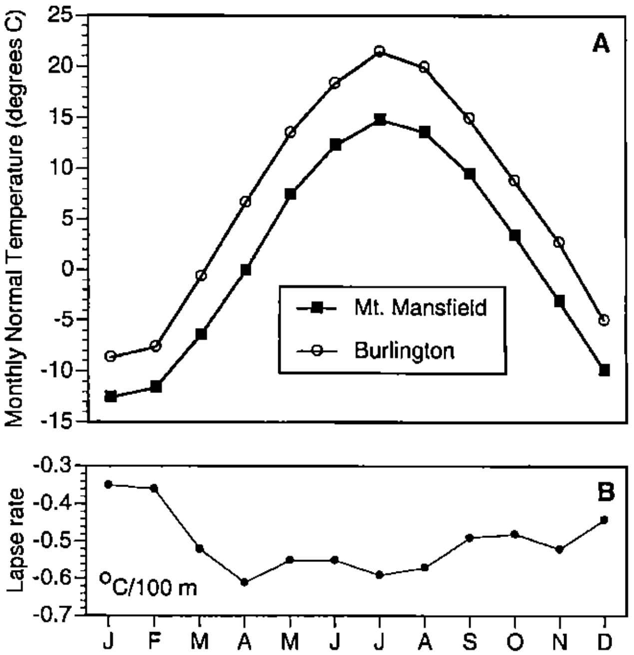


Figure 16: Monthly average temperatures and lapse rates for Mt. Mansfield weather station (1204 m asl) and Burlington weather station (101 m asl). Data from National Weather Service. A. Monthly average temperatures. B. Effective monthly average lapse rates.

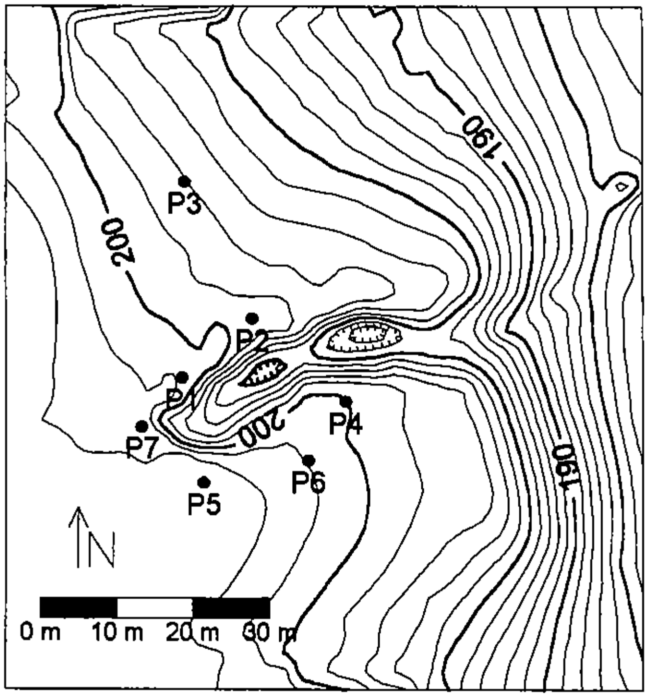


Figure 17: Map of gully and terrace immediately above active Stowe alluvial fan. Map produced using Pentax Total station and Trimble GPS 4400 RTK system by University of Vermont, 1998 Geohydrology class.

near the margin of glacial Lake Vermont at Vermont's premier attraction.

Stop 2-1 — Smugglers Notch

Smugglers Notch is a deep cleft that cuts across the main range of the Green Mountains just north of Mt. Mansfield. The extreme topographic relief, recent landslide scars, and large truck-sized blocks of rock that have fallen from the cliffs high overhead make this a much-visited site. Slope stability history and hazards in Smugglers Notch are discussed in Lee et al. (1994) and the 1983 slope failure is described in some detail by Baskerville et al. (1988).

The rocks exposed in the cliffs above Smugglers Notch are all schists belonging to both the Underhill and Hazen's Notch Formations. The foliation is defined largely by both muscovite and chlorite and, along the main range of the Green Mountains, has been folded to form the Green Mountain Anticlinorium. In Smuggler's Notch, the layering in the cliffs high overhead is almost horizontal and the Underhill Formation structurally overlies the Hazen's Notch Formation.

Smuggler's Notch does not make a straight knife-like cut across the Green Mountains, but instead is segmented into three relatively straight sections that are probably controlled by joints. Although the valley fill hides the bedrock along the floor of the notch and the structures contained therein, joint sets in the adjacent cliffs that are parallel to the valley segments are visible on aerial photographs and were measured by Lee et al. (1994).

Rock falls and debris flows are relatively frequent events in Smugglers Notch, many of which have been documented in the last 150 years (Lee et al., 1994). No bedrock is exposed anywhere along the floor of the notch. Most of the larger material transported to the bottom of the notch remains there and consequently the floor of the notch is gaining elevation as the sides widen. Structural controls that determine the dimensions of the cliff-loosened blocks include the horizontal foliation, the position of the relatively stronger Underhill Formation rocks above the weaker Hazen's Notch Formation rocks, and the joint sets.

We will observe the debris slide that occurred on July 13, 1983 and that is described by Baskerville et al. (1988). The landslide began at about 7 a.m. when a large block of rock (~10.4 × 106 kg) that cantilevered over the valley broke loose and fell onto the talus slope at the base of the cliff. The fall initiated a debris slide along the talus slope and material moved as far as the road (Baskerville et al., 1988). The rock fall occurred on a clear, sunny, midsummer morning and no rain had fallen for several days. Baskerville et al. (1988) suggest that the rock failure was most likely due to thermal expansion of the rock along a crack that had previously been extended by frost wedging.

Smugglers Notch Debris Flow

The debris flow deposit visible along Rt. 108 near the Cambridge/Stowe town line is one of several that occurred during the night of May 22, 1986 and are described by Lee et al. (1994). This particular flow originated in the gully extending up the east side of the valley below Spruce Peak and incorporates colluvium as well as trees and soil; approximately 250,000 m³ of material (Lee et al., 1994). An intense rainfall apparently precipitated this and other debris flows that evening, loosening colluvium and organic debris that had accumulated in the chute. At present, this debris flow chute is almost barren of colluvium and will take some time to once again accumulate sufficient debris so as to present a hazard.

Stop 2-3 — Mt. Mansfield

Mt. Mansfield, at 1340 m (4,393') is the highest point in Vermont. It is one of only five peaks in Vermont that exceed 1220 m (4,000') in elevation, the others being Camels Hump at 1248 m (4,093'), Killington Peak at 1293 m (4,241'), Mt. Ellen at 1245 m (4,083'), and Mt. Abraham at 1221 m (4,006'). Mt. Mansfield's rocky summit exposes multiply deformed schist of the Underhill Formation (Christman and Secor, 1961) covered in places by a thin mantle of till. Although the rock has been eroded into streamlined forms, >12,000 years of post glacial weathering and erosion have removed striations except in areas recently exposed by human activity. Glacial grooves, striations, the orientation of streamlined features, and erratics derived from the Champlain Lowland can be used to show that ice flowed approximately NNW to SSE over the summit (Fig. 7).

Mt. Mansfield has a weather station at the summit where precipitation and temperature have been measured since 1955. In general, annual precipitation is well-correlated to the Burlington station and about twice as abundant (1880 mm/74 inches vs. 890 mm/ 35 inches) due to orographic lifting. Lapse rates vary seasonally based on Burlington observations made at 101 m asl (Figure 16). Winter inversions reduce effective

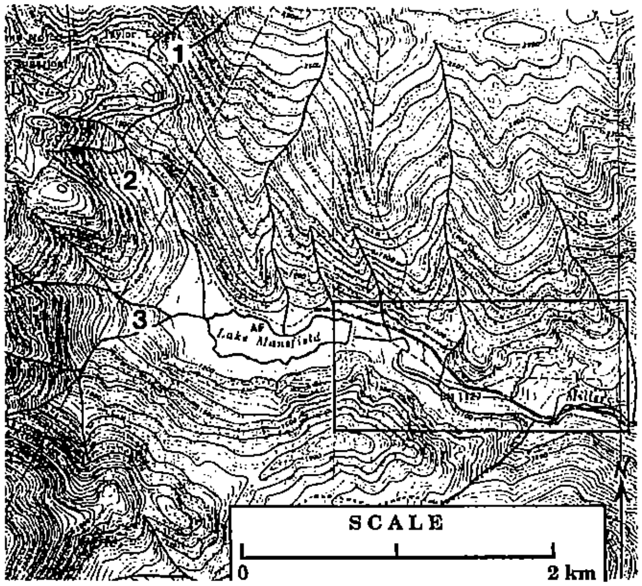


Figure 18: Portion of the Bolton Mountain 7.5' quadrangle map (1948, photorevised 1983, 1:24,000, 20 ft contours) showing the Miller Brook Valley from its headwaters in Nebraska Notch . The head of the Miller Brook Valley contains three cirque-like landforms (bowl-shaped with steep sides and a gently sloped floor): (1) the valley immediately east of Taylor Lodge, (2) the north-northwest-trending portion of the valley NNW of Lake Mansfield, and (3) the portion of the valley occupied by Lake Mansfield.

average lapse rates from about 0.6° C/100 m in late spring and summer to <0.4° C/100 m in January (Lipke and Pickard, 1993).

The relatively harsh climate on the top of Mt. Mansfield affects the vegetation. The summit area is dominated by a boreal assemblage of trees and tundra species typically found hundreds of kilometers farther north. Stunted spruce and fir are common as is paper birch. Wind-driven icing in wintertime controls treeline, which is lower on the west than the east side of the mountain. On and near the summit, it is not uncommon to see bog species such as sphagnum moss which thrive in nutrient-poor, acidic environments.

On a clear day the summit provides views extending over 100 km. To the east is the Stowe Valley and Worcester Range in the foreground, with the White Mountains of New Hampshire in the background. To the south is the spine of the Green Mountains and Camels Hump. To the west is the Champlain Lowland and beyond, the Adirondacks. To the north, are the lowlands of southern Canada and the Richelieu River, the outlet of Lake Champlain.

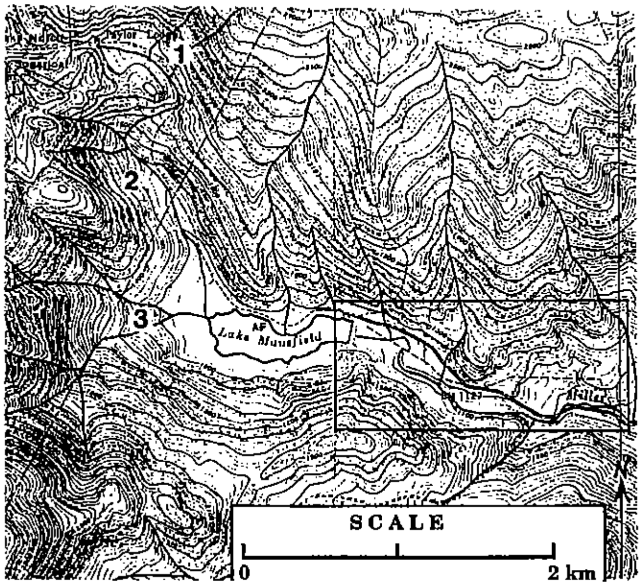


Figure 18: Portion of the Bolton Mountain 7.5' quadrangle map (1948, photorevised 1983, 1:24,000, 20 ft contours) showing the Miller Brook Valley from its headwaters in Nebraska Notch . The head of the Miller Brook Valley contains three cirque-like landforms (bowl-shaped with steep sides and a gently sloped floor): (1) the valley immediately east of Taylor Lodge, (2) the north-northwest-trending portion of the valley NNW of Lake Mansfield, and (3) the portion of the valley occupied by Lake Mansfield.

Stop 2-4 — Stowe Fan and Gully

West of Moscow, Vermont, is the most frequently active alluvial fan that we have identified so far in northwestern Vermont. The fan has a low gradient at the apex (4°) and the toe merges imperceptibly with the underlying fluvial terrace. The fan is active several times a year following heavy rain or snow melt events and sediment deposition on the fan appears to be solely by stream flow. We have repeatedly observed shallow (<10 cm) fan-head trenching. The fan is composed of reworked, fine-grained glacial-lacustrine sediments eroded from the terrace above. An adjacent gully indicates that till extends under the fine-grained sediments.

A long (~50 m) narrow (~6 m) gully supplies sediment to the fan (Fig. 17). Within the uncertainty of our calculations, the gully and the fan volume are similar (830 m^3 vs. 1100 m^3, respectively). There is no surface drainage in the gully. Sediment leaves the gully through a natural piping network below the gully bottom; the pipe daylights about half way down the terrace riser. These pipes may be created when blocks of cohesive wall material topple or rotate into the bottom of the gully. Dye tracing of pipe flow conducted during an extended dry period suggests unconstricted flow through the pipe. On the south side of the active gully is a relict gully, now apparently stable as indicated by the presence of mature trees.

This site illustrates the interdependence of groundwater flow and slope stability. Failure of the gully walls by toppling and rotation appears to occur when the water table and thus pore pressures along the gully walls are high. Nested piezometers indicate that both the active and now-stabilized gullies act as groundwater drains lowering the water table. The relatively low hydraulic conductivity of the glacial-lacustrine sediments (< 1.5 x 10^{3} cm s^{1}) results in large pore pressure gradients near the gully walls. The south side of the gully, where the water table is lowered by the proximity of a stabilized gully 25 m away, maintains a shallower face than the northern wall of the gully, where the water table is higher. Mass wasting on the south side occurs primarily by slumping, freeze thaw, and soil creep. The north side erodes primarily by a combination of toppling and rotational failure.

The fan and the gully appear to be quite young. A soil pit dug near the fan apex revealed an old road surface more than a meter below the current land surface. The road was relocated in the late 1960s or early 1970s, suggesting that most if not all deposition on the fan occurred within approximately the last 30 years. Just above the road surface, buried by fan sediments, was an automobile part confirming that the recent period of fan activity has lasted no longer than the last 20 to 30 years.

We are uncertain what initiated gully formation here. Neighbors believe that the gully formed in response to clearcutting and the construction of logging roads. However, analysis of aerial photographs from 1943 onward shows that the slopes above the gully were clear for many years before erosion began (Flemer, 1998). Furthermore, examination of the town records show that this area was first cleared for farming before 1856 (Flemer, 1998). Comparing the size of the currently active gully with that of the adjacent gullies, and considering the average rate of erosion and sediment deposition on the fan over the past 20 years, we suggest that it will take the better part of a century for the gully to reach the depth and size of its neighbors.

Stop 2-5 — Lake Mansfield and Miller Brook Deposits

For more than 25 years, workers have debated the presence of late Pleistocene alpine ice in the mountains of Vermont and elsewhere in New England (e.g. Goldthwaitt, 1916; see review by Waitt and Davis, 1988). Two lines of evidence have been used to argue for post-Laurentide alpine glaciation: the presence of cirque-like features and the existence of linear ridges of sediment, interpreted as moraines. Both of these landforms are well-exhibited near Lake Mansfield in the Miller Brook Valley, south of Stowe, Vermont.

The Miller Brook Valley trends ESE from a low point along the crest of the Green Mountains (Nebraska Notch, 576 m asl, 1890 ft asl) (Fig. 18). Bedrock along the Miller Brook Valley consists of medium grade schists belonging to both the Underhill and Hazens Notch Formations (Christman and Secor, 1961). The dominant foliation in these rocks strikes N-S and dips moderately to steeply east. Resistant rock units within these formations control the orientation of tributary brooks to Miller Brook as well as the orientation of Miller Brook itself upstream of Lake Mansfield. The ESE trend of the Miller Brook Valley, and many other river and stream valleys in the Green Mountains, is probably controlled by similarly oriented zones of joints or brittle strike-slip faults, traces of which are readily visible on aerial photographs.

Beginning at Nebraska Notch, Miller Brook flows through three steep-sided, bowl-shaped segments of the valley (labeled "1, 2, and 3" on Figure 18), interpreted by some as cirques (Wagner, 1971). Lake Mansfield occupies the lowest of these and is artificially dammed. A large alluvial fan protrudes into the lake from the north. The stream feeding this fan is currently incised into the fan. In the area mapped below Lake Mansfield, Miller Brook lies entirely within surficial materials, stepping down through progressively lower lacustrine and fluvial terraces until entering the Little River just upstream from the Waterbury Reservoir. Downstream from the Lake Mansfield dam, the valley contains several distinct ridges (with up to 30 m of relief) lying within or adjacent to fluvial landforms and older lacustrine terraces.

Wagner (1970) first identified these ridges and interpreted them as moraines produced by alpine ice retreating up the Miller Brook Valley. Wagner's paper stimulated considerable discussion, mostly centered on whether or not the ridges were produced by a waning tongue of the Laurentide icesheet or by a local glacier (Stewart, 1971; Wagner, 1971; Ackerly, 1989; Waitt and Davis, 1988; Loso et al., in press) and raised again the question of whether or not alpine glaciers existed in New England following retreat of the Laurentide icesheet. We re-evaluate Wagner's (1970) hypothesis, using newly constructed maps and soil pit data.

Detailed Mapping

Interpreting landforms in the Miller Brook Valley has always been confounded by the lack of adequate maps. The ridges that Wagner (1970) identified as moraines are crudely located on his map. During the fall of 1995, the area immediately downstream of the Lake Mansfield dam was mapped at a scale of 1:2,500 by two students, M. Loso and H. Schwartz, using a Pentax Total Station (Loso et al., in press). The mapping area was extended downstream by Wright in the fall of 1996 (Wright et al., 1997), initially using tape and compass methods (1:2,000), with parts later surveyed for elevation control using the Total Station. A copy of the Wright et al. (1997) map will be supplied during the fieldtrip to participants.

Description of the Ridges

Ridges of surficial material in the Miller Brook Valley extend from ~150 m west of the Lake Mansfield dam, across the entire mapped area, and at least another two km down-valley. More detailed discussion of the ridges is presented in Wright et al. (1997). For clarity, individual ridges have been designated with letters, e.g. Ridge C, on the map The ridges generally follow sinuous pathways, both along the side and across the middle of the valley and are cut by both ephemeral streams and Miller Brook. In places the ridges appear to bifurcate or to show cross-cutting relationships with one another (e.g. Ridges C and D, and Ridges F and G). The ridges are both sharp-crested (Ridges A, B, the eastern end of C, E, F, and G) and rounded (most of Ridge C and Ridge D), applying terminology employed by Shreve (1985). Ridge tops range from 4 m to more than 25 m above adjacent fluvial or lacustrine terraces. In general, the crests of Ridges A, B, and C gradually drop in elevation down valley from Lake Mansfield. An exception to this is the abrupt 14 m rise (to 362 m) and fall along a 100 m reach SW of Middle Pond.

Several observations indicate that the ridges are primary constructional landforms and not erosional remnants: (1) Ridge A surrounds a small bog ("B" on the map, adjacent to Soil Pit T2H). Sperling et al. (1989) retrieved a core from this bog from which they radiocarbon dated sediment recovered from between 2.75 and 2.85 m depth, 9,280±235 ^{14}C y BP. This age implies that the bog has been separated from the valley bottom by the intervening ridge for over 9,000 ^{14}C y, within 3,000 ^{14}C y after ice retreat (Li, 1996). (2) Soil pits adjacent to Ridges A (T3R), C, and D clearly show fine to medium sand onlapping the coarse sand, gravel, and rounded cobbles comprising the adjacent ridges, indicating that the ridges were partially buried by fluvial and lacustrine sediments after their formation. (3) A small alluvial fan, fed by an ephemeral stream 100 m NW of the previously described bog, has partially buried Ridge A. The active Holocene history of fans elsewhere in the region, suggest that this fan too has been actively depositing material against the ridge throughout the Holocene.

Ridges A, B, and C are the "classic" Lake Mansfield ridges that Wagner (1970) interpreted as moraines. Ridge A begins just beyond a broad area of hummocky relief including one well-defined, kettle-like closed depression where the trails intersect, south of Soil Pit RTP. This ridge, interpreted by Wagner (1970) as a lateral moraine, parallels the south side of the valley. No correlative ridge exists along the north side of the valley. Approximately 600 m below the Lake Mansfield dam, the ridge turns abruptly NE (Ridge B) and crosses the valley where it is cut by Miller Brook. Wagner (1970) interpreted this segment as an end moraine. However, Ridge B does not curve up-valley to meet a lateral moraine along the north side of the valley; rather, it strikes straight across the valley, turns slightly and continues down the middle of the valley (Ridge C; upon which the road is built) until it is cut again by Miller Brook, just east of the Mill Pond dam.

Soil pits were dug either along or adjacent to Ridges A and B as part of the studies presented by Loso et al. (in press) and Wright et al. (1997). With few exceptions, most of the pits were excavated to 1 m depth. The near-surface materials comprising Ridges A, B, and C are quite similar to one another and texturally bear attributes of both till and fluvial sediments. Soil Pit C-2, along the crest of Ridge C, exposes 0.5 m of poorly sorted, poorly to well rounded coarse sand, gravel, and cobble clasts in a fine sand matrix. These materials overlie clean gravel and coarse sand extending to the bottom of the pit (0.7 m). Soil Pit T1M, along the crest of Ridge B, similarly contains a poorly sorted mixture of moderately to well rounded coarse sand, pebbles, and cobbles in a medium to fine sand matrix. A recent soil slip along the NE side of Ridge B (~5 m downstream from Soil Pit T1R,) reveals clean, moderately sorted and rounded, coarse sand and gravel just above water level that is overlain by at least 3 m of material similar to that exposed along the crest of the ridge. The gravels occurring in both Ridges B and C are clearly fluvial and the overlying material in both ridges contains abundant rounded clasts, also indicative of fluvial processes.

Based on both their map pattern and the fluvial sediments within them, we interpret these ridges to be segments of an esker, and not moraines. The poorly sorted veneer of sediment overlying the gravels likely originated from the low discharge of the stream occupying the esker tunnel towards the end of its history. The rounding of most clasts indicates at least some transport of the melt-out till. Soil pit C-1 on Ridge C contains coarse diamict including a few faceted, striated cobbles. It seems likely that this diamict melted out of the ice shortly before the ice tunnel was abandoned and that these clasts were never subjected to significant abrasion.

Discussion

The direction of ice flow in this part of Vermont was generally from NNW to SSE, across the N–S trending ridge of the Green Mountains (Stewart and MacClintock, 1970; Christman and Secor, 1961; Larsen, 1987a; Ackerly and Larsen, 1987), as evidenced by grooves and striations and Champlain Lowland erratics preserved along the crest of the Green Mountains. Once the elevation of the ice surface dropped below the N–S ridge of the Green Mountains (Fig.1), continued ice flow from the NNW was funneled through, from north to south, the Lamoille River Valley (elev. 150 m, 490 ft), Smuggler's Notch (elev. 670 m, 2200 ft), Nebraska Notch (elev. 576 m, 1890 ft), and the Winooski River Valley (elevation 103 m, 340 ft) . Cores taken from Sterling Pond (1.7 km east of Smuggler's Notch, elev. 915 m, 3,000 ft) indicate that the ice was below this elevation at least by 12,700 ^{14}C y BP (Li, 1996).

Nebraska Notch is the lowest gap through the Green Mountains between the Lamoille River Valley to the north and the Winooski River valley to the south. Therefore, the Miller Brook Valley must have contained Laurentide ice well after ice had melted from the eastern flanks of the Green Mountains immediately to the north and south. Once the ice elevation dropped below the elevation of Nebraska Notch, ice supply to the Miller Brook valley was shut off and the ice in the valley stagnated soon thereafter. The esker documented in this report probably formed immediately before and during this stagnant ice stage. While not mapped in detail, reconnaissance shows the esker continuing at least 2.6 km down-valley from the Lake Mansfield dam, the crest dropping in elevation from 352 m just below the Lake Mansfield dam to 299 m at its most eastern end. Ice-contact deformation of the lacustrine sediments underlying the 341.5-m terrace implies that this lake occupied the Miller Brook Valley during the final stages of stagnant ice melting. It is unclear whether or not this lake was ever continuous with Lake Winooski, although the lake producing the lower 330-m terrace is low enough to make its correlation with Lake Winooski likely.

Loso et al. (in press) have determined that the floor of the Miller Brook Valley is too low to maintain a valley glacier after the retreat of the Laurentide icesheet. Given our current understanding of climatic cycling during the Quaternary where each glacial maximum is preceded by gradual cooling and followed by rapid warming (e.g. Dansgaard et al., 1993), it is likely that the headwaters of the Miller Brook Valley (if they are a cirque) were shaped during the onset of different glacial episodes, before being overridden by continental ice. Waitt and Davis (1988) have pointed out that alpine glacial landforms can be preserved, despite subsequent cover by continental ice. In this light, the cirque-like landforms at the valley head, could have been filled with alpine ice before the onset of each continental glaciation, whereas the ridges of surficial material were formed as eskers within an ice tongue of the thinning Laurentide icesheet.

Stop 2-6 — Ben and Jerry's Factory

The Ben and Jerry's factory in Waterbury is currently the most popular tourist attraction in Vermont. The factory is located at an elevation of about 180 m, an area that would have been on the bottom of high level lakes in the Winooski Valley but near the shoreline as the icesheet retreated, establishing Lake Vermont. By Champlain Sea time, the factory would have stood on dry land.

ACKNOWLEDGMENTS

We thank the Moultroup and Aldrich families of Huntington and the Lake Mansfield Trout Club for allowing us to dig into the landforms on their properties. Thanks are also extended to Lori Barg for assisting with surveying in the Mill Pond area and Laura Mallard for editing this field guide.

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