LATE PLEISTOCENE–HOLOCENE HISTORY: HUNTINGTON RIVER AND MILLER BROOK VALLEYS, NORTHERN VERMONT

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INTRODUCTION

The areas visited on this field trip are within the Huntington River valley, draining the western slopes of the Green Mountains immediately south of the Winooski River, and the Miller Brook valley, a much smaller, east-flowing drainage basin north of the Winooski River (Fig. 1). Our objectives are to present the results of recently completed and ongoing research at the University of Vermont centered on better understanding the Late Pleistocene and Holocene history of these valleys. Parts of the material presented in this article appear in University of Vermont theses by Zehfuss (1996), Whalen (1997), Church (1997), and Bryan (1995). A comprehensive summary of regional Holocene findings appears in Bierman and others (1997) and a progress report on work in the Miller Brook valley was published by Loso and others (1997).

In the following paper we first outline the Late Pleistocene history of the Miller Brook valley inferred from our current work in the context of earlier studies of the valley. Central to this section is the reinterpretation of glacial landforms previously identified as moraines. In particular, we outline the evidence for an ice contact environment in the valley at the time most of the surficial materials were deposited. We next present a summary of the glacial lake history and later fluvial history of parts of north-central Vermont derived from recent surveying. We follow this with a review of information gleaned from numerous alluvial fans, built on both lacustrine and fluvial terraces, whose history spans the entire Holocene. We present detailed maps, logs of trenches, and ^{14}C dates as the primary data from which we base our interpretations.

The field sites described in the following text are found on the "Huntington" and "Bolton Mountain" 7.5' U.S.G.S. Quadrangle maps. Parts of these maps are reproduced as part of this field guide. Most of the field stops described herein occur on private property and permission needs to be secured before venturing onto these field sites. Contacts are listed in the Road Log.

ICE-CONTACT ENVIRONMENT: MILLER BROOK VALLEY

Stephen F. Wright

Introduction

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. 576 m, 1890 ft) to its confluence with the Little River (Elev. 184 m, 605 ft), (Figs. 1 and 2). The bedrock geology of the area is described by Christman and Secor (1961) although more recent and detailed mapping has been completed and will be incorporated into the new State Geologic map. Rocks along the Miller Brook valley consist of medium grade schists belonging to both the Underhill and Hazens Notch formations. The dominant foliation in these rocks (S_2) strikes N–S and dips moderately to steeply east. Resistant rock units within these formations "V" downstream and control the orientation of tributary brooks to Miller Brook as well as the orientation of Miller Brook itself upstream of Lake Mansfield (Fig. 3). 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 (see Christman and Secor, 1961, for compilations of brittle structures in the area).

Beginning at Nebraska Notch, Miller Brook flows through three steep-sided, bowl-shaped segments of the valley (labeled "1, 2, and 3" on Fig. 3). Lake Mansfield occupies the lowest of these flat-bottomed, steep sided valley segments 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. Below Lake Mansfield in the area mapped, 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 lying within or adjacent to fluvial landforms and older lacustrine terraces and standing with up to 30 m of relief. The origin of these ridges and the surrounding sediments is the subject of this paper.

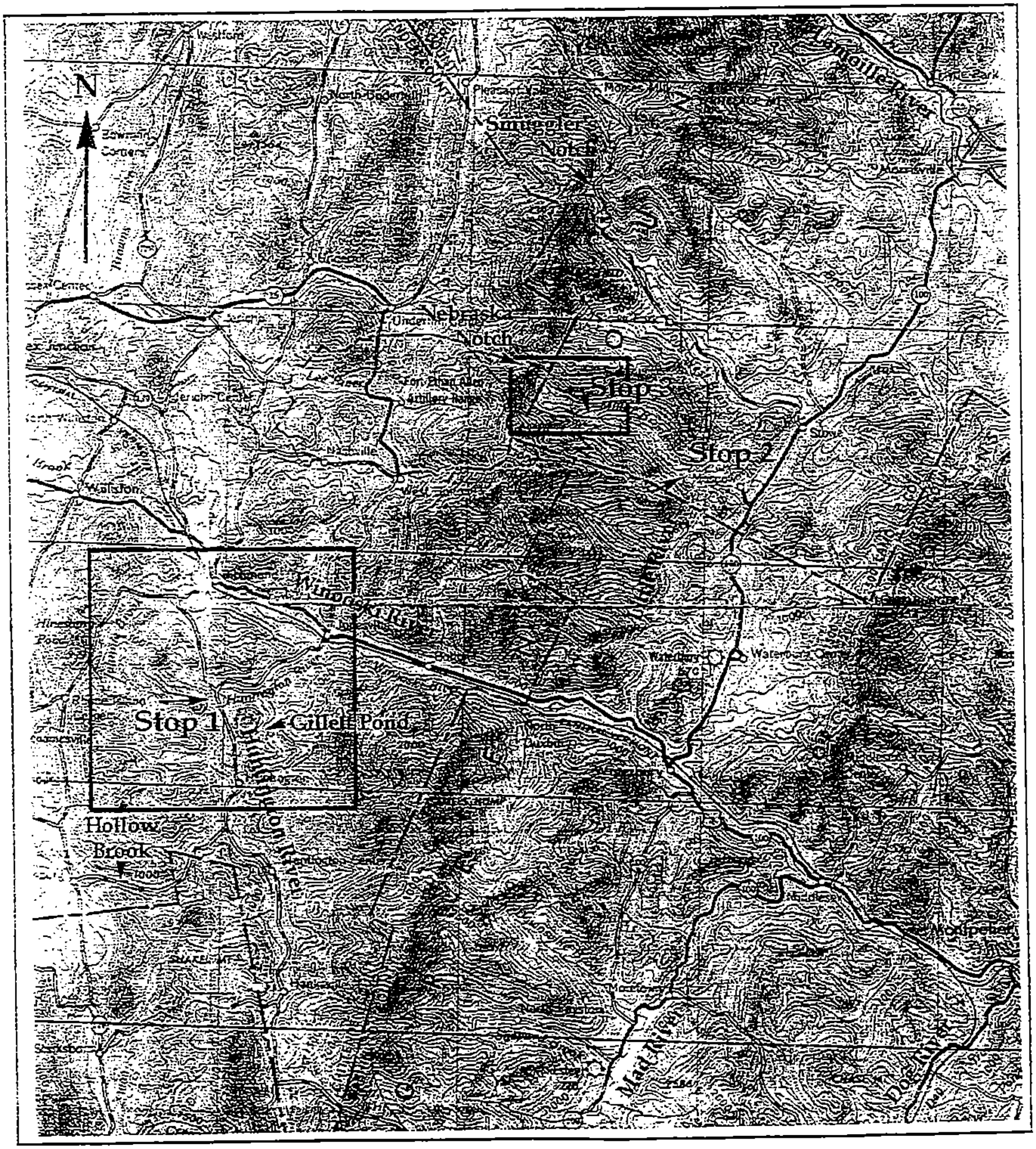


Figure 1: Shaded relief map of north-central Vermont showing the location of field stops discussed in this guide. Drainage basins discussed in the text are indicated on the map. Ice flow direction was generally to the SSE. across the N–S trending ridge of the Green Mountains based on striae preserved along the Green Mountain Ridge (Stewart and MacClintock, 1970). Note that Interstate 89 is not shown traversing through the Winooski Valley on this map. Grid outlines 10 km squares. Upper box outlines Fig. 3 and lower box outlines Fig. 12.

Wagner (1970) first identified several ridges of surficial material in the Miller Brook valley and interpreted these to be moraines produced by a tongue of ice retreating up the Miller Brook Valley. He further conjectured that the Miller Brook moraine (and other moraines he identified in northern Vermont) were not produced by the waning Laurentide ice sheet, but by independent alpine glaciers that occupied the valley subsequent to ice-sheet retreat. Wagner's paper stimulated considerable discussion, mostly centered on whether or not the ridges were produced by a waning tongue of the Laurentide ice sheet or by a local glacier (Stewart, 1971; MacClintock, 1971; Wagner, 1971; Ackerly, 1989; Waitt and Davis, 1988) and raised again the question of whether or not alpine glaciers existed in New England following retreat of the Laurentide ice sheet (e.g. Goldthwait, 1916; see review by Waitt and Davis, 1988). Wagner's hypothesis is reevaluated below, using newly constructed maps and soil pit data. At least a portion of the area visited on this field trip was examined on an earlier NEIGC field trip led by Wagner (1972).

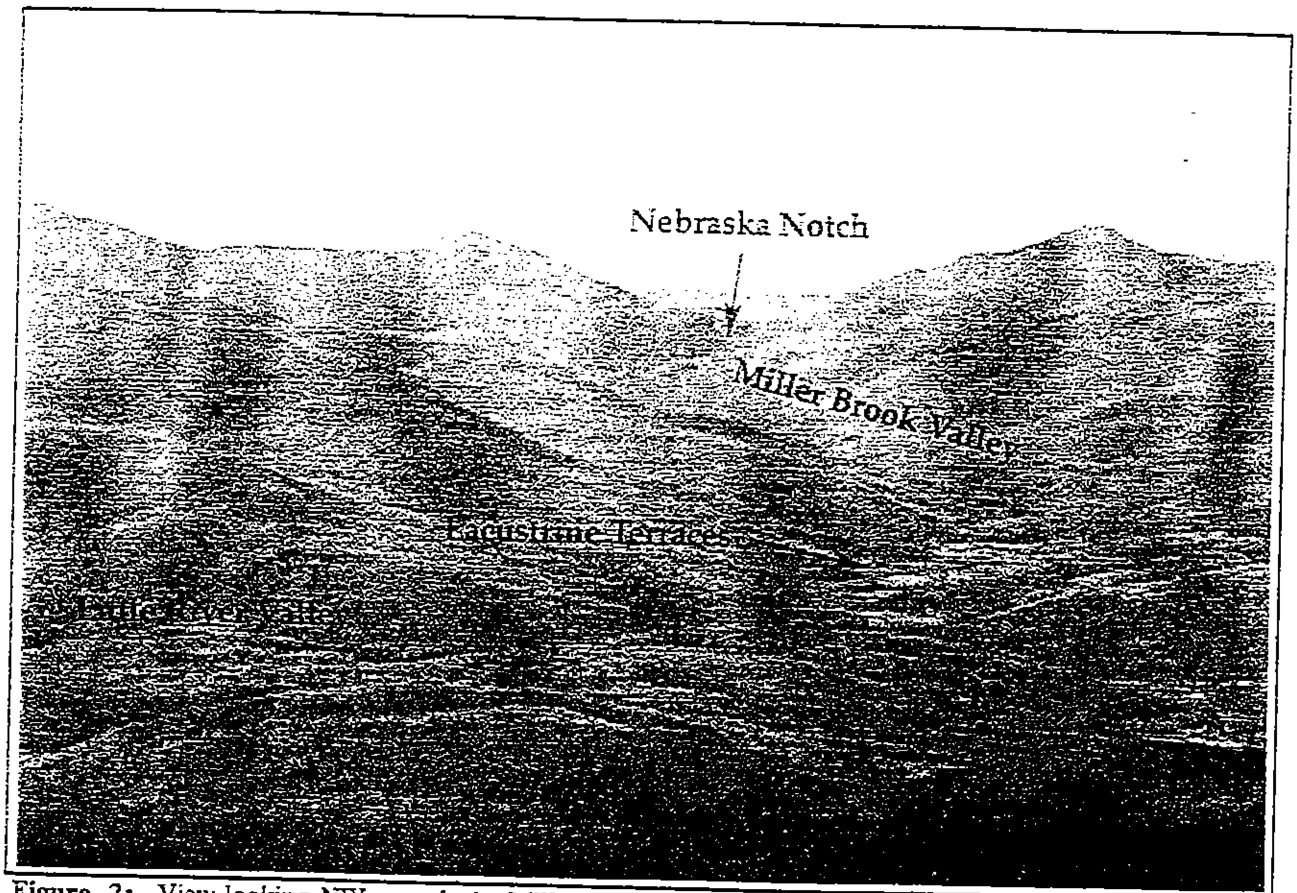


Figure 2: View looking NW towards the Miller Brook valley and Nebraska Notch (elevation 576 m, 1890 ft) from Hunger Mountain. Nebraska Notch is the lowest pass through the mountains between the Lamoille River Valley to the north and the Winooski River Valley to the south. Dewey Mountain (elev. 1024 m, 3360 ft) lies to the north of the Notch and Mount Clark to the south (elev. 902 m, 2960 ft). Small open meadows along the lower part of the Miller Brook valley are on a flight of terraces ranging from 256 to 198 m (840 to 650 ft) corresponding to Lakes Mansfield I and II and the Quaker Springs stage of Lake Vermont.

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 and his brief descriptions are insufficient to clearly locate their position or morphology. Only portions of the ridges appear on the U.S.G.S. topographic map (Bolton Mountain Quadrangle) and the position of Miller Brook and its principal tributary (unnamed) from the SW are incorrectly shown. 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 Total Station. This map and preliminary descriptions and interpretations of soil pits constructed by P. Bierman's geomorphology class are presented in Loso and others (1997). The mapping area was continued downstream by the author in the fall of 1996, initially using tape and compass methods (1:2,000) with parts later surveyed for elevation

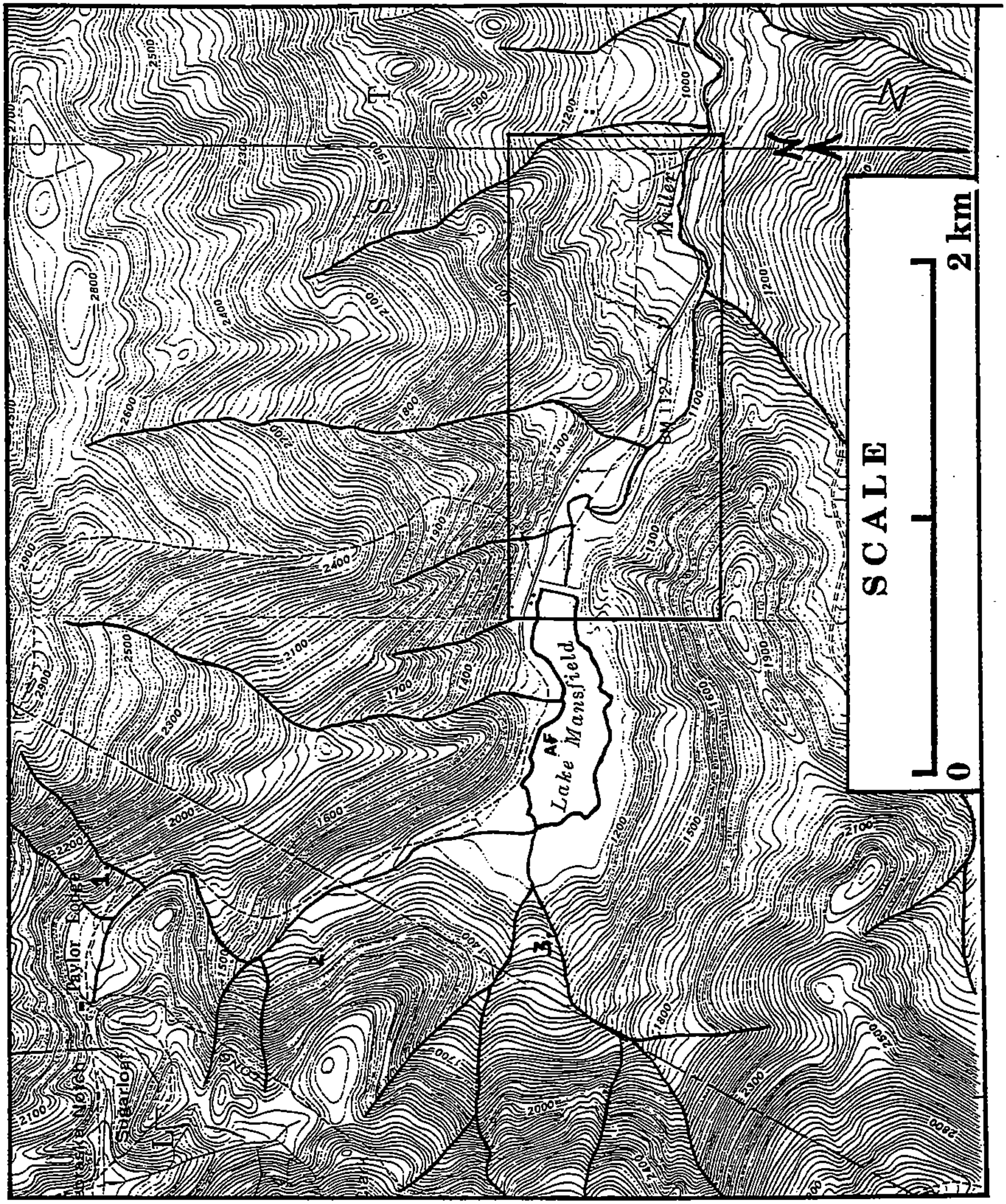


Figure 3: Portion of the Bolton Mountain 7.5' Quadrangle Map (20 ft contours) showing the Miller Brook Valley from its headwaters in Nebraska Notch . The head of the Miller Brook Valley contains three cirque-like landforms (bowl-shaped with steep sides and a gently sloped floor): **(1)** the valley immediately east of Taylor Lodge, **(2)** the NNW-trending portion of the valley NNW of Lake Mansfield, and **(3)** the portion of the valley occupied by Lake Mansfield. Box outlines detailed map of Surficial Landforms (Fig. 4). **AF** = Alluvial Fan

control using the Total Station. The map presented in this field guide is a compilation of these two maps set within an enlarged portion of the U.S.G.S. base map (Fig. 4). The new mapping uses 2 meter contours tied to a benchmark at 1127 ft, whereas the surrounding U.S.G.S. topographic map uses 20 ft contours.

Description of the Landforms

Ridges of surficial material in the Miller Brook valley extend from ~150 m west of the Lake Mansfield dam (not mapped on Fig. 4), across the entire mapped area (Fig. 4), and reconnaissance work indicates that they extend at least another kilometer down-valley. For clarity, individual ridges have been designated with letters, e.g. Ridge C, on Figure 4. 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 they appear to bifurcate or to show cross-cutting relationships with one another (e.g. Ridges C and D, and Ridges F and G, Fig. 4). 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) and show relief ranging from 4 to in excess of 25 m above adjacent fluvial or lacustrine terraces. In general, the crest of Ridges A, B, and C gradually diminishes from 352 m just below the Lake Mansfield dam to 342 m NE of the Mill Pond dam. An exception to this is the abrupt 14 m rise (to 362 m) and fall along a 100 m reach SW of the Middle Pond (Fig. 4).

That the ridges are primary constructional landforms and not erosional remnants is evidenced by several observations: (1) Ridge A surrounds a small bog ("B" in Fig. 4, adjacent to Soil Pit T2H). Sperling and others (1989) retrieved a core from this bog and present a ^{14}C date of 9,280 ± 235 yBP (years before present) from sediment recovered between 2.75 and 2.85 m depth, implying that the bog has been separated from the valley bottom by the intervening ridge for over 9,000 ^{14}C years, within 3,000 ^{14}C years since ice retreat (Lin Li, 1996). (2) Soil pits adjacent to Ridges A (T3R), C, and D clearly show fine to medium grained 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 (Fig. 5). (3) A small alluvial fan, fed by an ephemeral stream 100 m NW of the previously described bog, has partially buried Ridge A (Fig. 4). By inference to the active Holocene history of fans elsewhere in the region (see later sections of this paper) we suggest that this fan too has been actively depositing material against the ridge throughout the Holocene.

Ridges A, B, and C

The "classic" Lake Mansfield ridges, the ones Wagner (1970) describes as forming during a later phase of glaciation (his Phase II), are labeled Ridges A and B on Figure 4. 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) to be 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, Fig. 4) and crosses the valley where it is cut by Miller Brook. This is the segment interpreted to be an end moraine by Wagner (1970). Based on the map pattern shown in Figure 4, Ridge B does not curve up-valley to meet an unidentified lateral moraine along the north side of the valley, but instead strikes straight across the valley, turns slightly and continues down the middle of the valley (Ridge C; the road is built upon it) until it is cut again by Miller Brook, just east of the Mill Pond dam.

Ten soil pits were dug either along or adjacent to Ridges A and B (e.g. T2M, Fig. 4) as part of the study presented by Loso and others (1997). Several of the pits considered most important to this study were reopened and are reinterpreted here together with observations from new pits (e.g. C-2, Fig. 4). With few exceptions, most of the pits were excavated to approximately 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 (Fig. 4), 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 (Fig. 4), 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 in Soil Pit T1M 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, I interpret these ridges to be segments of an esker and not moraines. The poorly sorted veneer of sediment overlying the gravels has likely originated from the low discharge of the stream occupying the esker tunnel towards

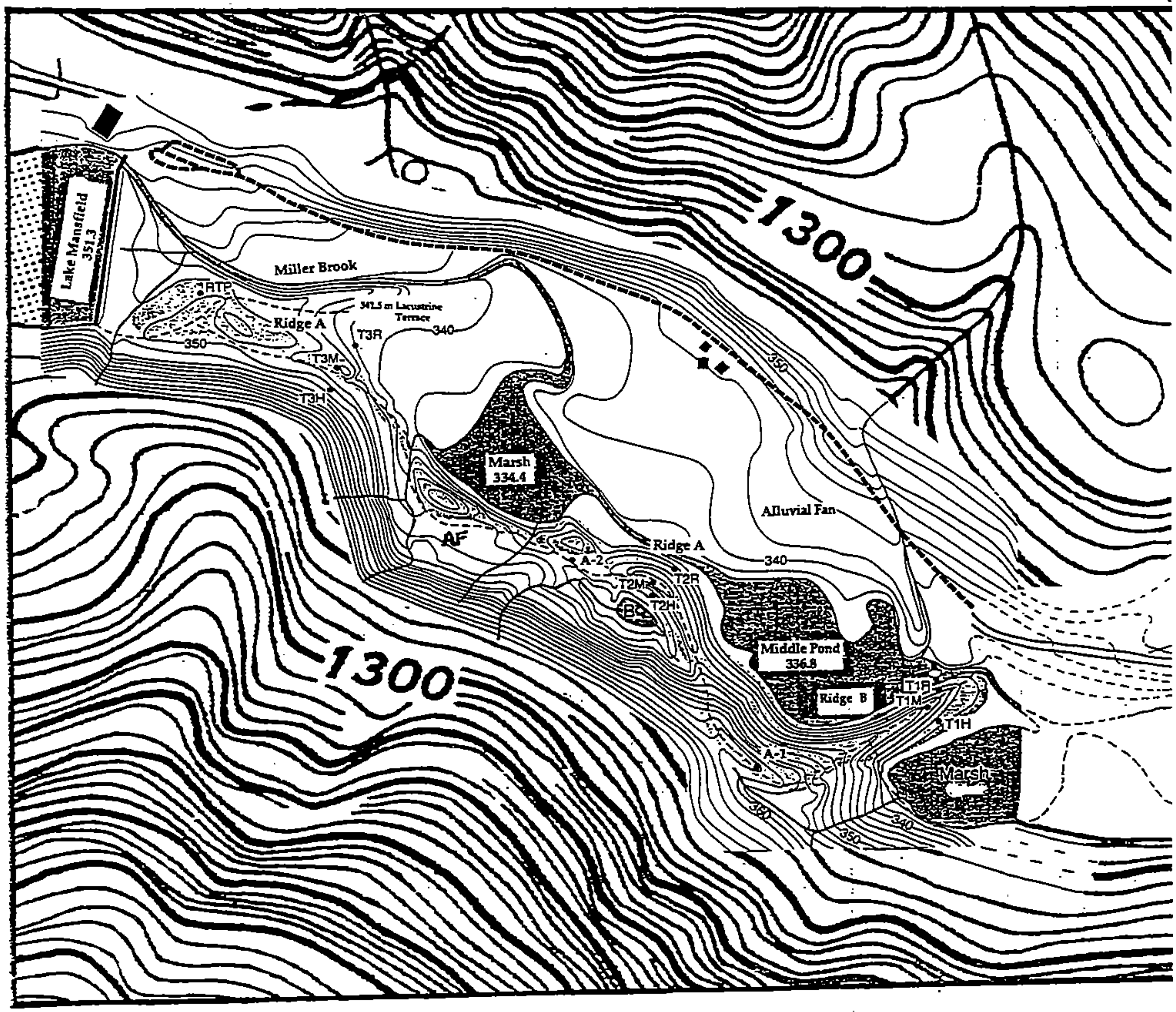


Figure 4: Detailed map of surficial landforms in the upper portion of the Miller Brook Valley (see Fig. 3 for exact location). Western half of map is by M. Loso and H. Schwartz (Loso et al., 1997) and uses 2 m contours. Eastern half of map is by S. Wright and that part surveyed also uses 2 m contours, dashed where inferred. Both maps are placed within an enlarged portion of the Bolton Mountain 7.5' Quadrangle map. Ridges are labeled with letters, e.g. Ridge A, and extend across the entire mapped area. AF = Alluvial Fan.

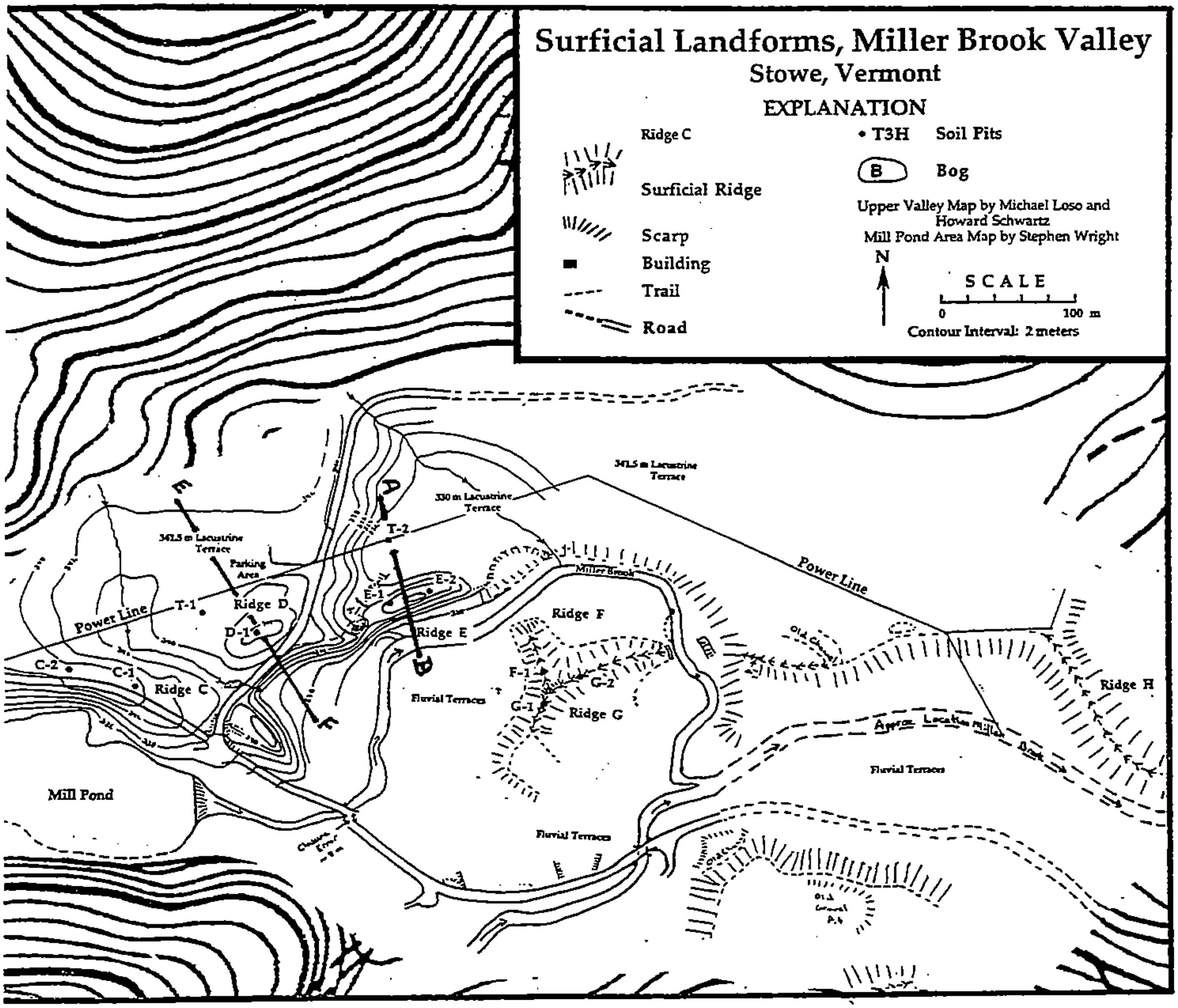


Figure 4: Continued from previous page.

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 (Fig. 4) 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 the clasts never subjected to significant abrasion.

The transition from Ridge B to Ridge A is marked by an abrupt change in orientation from N60E to N45W, rapid ascent to the highest elevation of any of the mapped ridges (362 m, Fig. 4), and the presence of numerous large, mostly angular boulders, many exceeding 2 m in diameter. Large boulders are also common along segments of Ridge A opposite the "Marsh" (at 334.4 m elev.) and in the vicinity of Soil Pit T3M, but much less common along the segment opposite the Bog ("B" on Fig. 4). Boulders of similar size and frequency are rare along the other ridges shown on Figure 4, all of which occupy positions considerably removed from the sides of the valley. Soil Pits along the crest of Ridge A (A-1, T2M, and T3M) all display unsorted, unbedded coarse sand, gravel, cobbles, and boulders in a medium to fine sand matrix with minor silt (see grain size distribution curves in Loso et al., 1997). Pebbles and cobbles excavated in A-1 and reexcavated in T3M are nearly all moderately rounded (T2M was not reexcavated). Soil textures in these pits are similar to the soil pit (T1M) on the crest of Ridge B, except for the large, usually angular boulders mentioned earlier. These materials differ from the colluvium along the adjacent hillside in that most of the clasts larger than 1 mm are rounded as opposed to angular or subangular. In the top meter of Ridge A none of the pits revealed layers of clean, coarse sand and gravel similar to those observed in Ridges B and C and it is therefore more difficult to claim that fluvial processes are responsible for the deposition of these materials. Nevertheless, the abundant rounded clasts, paucity of silt and clay, and textural similarity to the surficial materials overlying fluvial gravels in Ridges B and C are at least consistent with their deposition in a subglacial stream flowing along the side of the valley. Given the mapped continuity of Ridge A with Ridges B and C, I similarly interpret Ridge A to be an up-valley segment of the same esker. The abundant large boulders in and on it were most likely derived from materials sliding down the adjacent hillside onto the ice surface and later being incorporated into the esker as the tunnel ceiling was melted by the subglacial stream.

Ridges D, E, F, G, and H

Similar to Ridge C, described in the previous section, Ridges D, E, F, G, and H all lie in the central portion of the Miller Brook valley, well away from the valley sides (Fig. 4). Ridge D is a rounded, elongate hill rising in low relief relative (maximum elevation 345.3 m) to the surrounding terrace at 340.5 m elevation. It is oriented approximately N60E and may be continuous with Ridge E. Ridge E is sharp-crested, rises to 338.4 m and shows sharp relief with the 330 m terrace to the north and the <324 m fluvial terraces to the south. Soil pits in these ridges (D-1, E-1, E-2; Fig. 4) all contain clean, well rounded gravel and cobbles in a coarse sand matrix. Bedding was observed in Pit E-1. The sediments in these ridges are clearly fluvial in origin and I therefore interpret Ridges D and E to be segments of an esker. Ridge C appears to cross-cut Ridge D, implying that the former is younger, but it is also possible that the esker bifurcated at this point (Fig. 4).

Ridges F and G are the most striking in the valley forming sharp crested ridges rising at least 25 m above the fluvial terraces that surround them (Fig. 4). Ridge F may be the continuation of Ridge E as it arches across Miller Brook. Ridge F is clearly cross-cut by Ridge G, which rises approximately 5 m above Ridge F at their intersection. Ridge G strikes N15E at its southern end, but swings almost due east where it is cut by Miller Brook. Ridge H on the opposite side of Miller Brook is the continuation of Ridge G, rising to the 341.5 terrace level, where it is completely buried, and then reappears at the eastern edge of the map where it has been excavated by stream erosion (Fig. 4). Soil pits in these ridges (F-1, G-1, G-2) all contain unsorted, moderately rounded cobbles and pebbles in a fine to coarse sand and gravel matrix. Large (up to 1.5 m), relatively unrounded boulders are also incorporated into the ridges. The texture of surficial materials in these pits is very similar to that in the near-surface parts of Ridges B and C (see earlier descriptions). None of the soil pits exceeded 1 m depth and none exposed any clean gravel similar to that observed at the surface in Ridges D and E. Based on the abundance of rounded clasts in these ridges and their sinuous, mid-valley form, I also interpret these ridges to be eskers. Furthermore, it is likely that the easternmost end of Ridge C once connected with the southern end of Ridge G and the intervening esker has been removed by Miller Brook. Ridges D, E, and F may document a former esker tunnel that was abandoned in favor of the tunnel occupied by the eastern end of Ridge C and the southern end of Ridge G.

Lacustrine Terraces

Two distinct terraces exist in the map area, the upper one at ~341.5 m (~1,120 ft) and the lower one at ~330 m (~1,082 ft). Elevations are taken from the middle of these terraces which show ±0.4 m of minor relief. Materials exposed in soil Pits T3R, A-2, T-1, and T-2, excavated in these terraces, are described here as well as materials exposed in recent road-cuts in the vicinity of the Mill Pond dam (Fig. 4). Two meters of well sorted and bedded fine sand and silt with two thin beds (2–3 cm) of coarse sand onlap the coarse sediments of Ridge A in Soil Pit T3R, located adjacent to Ridge A, on the 341.5 m terrace (0.4 m of colluvium overlie the terrace deposits, Loso et al., 1997). One of the coarse sand layers contains load casts down into the underlying silts. Bedding strikes to 331 and dips 29° NE, parallel to the slope of the buried ridge. Some of the bedding has slumped to the NE, down the slope of the buried ridge. Loso and others (1997) interpret these sediments to be lacustrine, deposited in a shallow area of a lake whose minimum elevation was 342 m. Soil Pit A-2, dug along Ridge A, also reveals medium to fine sand with isolated, matrix supported, moderately rounded pebbles and cobbles. This pit is at the same elevation as Pit T3R and these sediments are probably also lacustrine in origin and onlap Ridge A.

Two topographic profiles (A–B and E–F, Fig. 4) show the terraces adjacent to Ridges D and E and the materials exposed in nearby soil pits and auger holes (Fig. 5). Soil Pit T-1, located on the 341.5 m terrace 40 m west of Ridge D (Fig. 4), was extended to 3.7 m depth using a bucket auger. The upper 1.4 m consists of clean, moderately sorted and well rounded, medium to coarse sand and gravel. The remainder of the auger hole (1.4 to 3.7 m) exposed only fine sand with rare layers of medium to coarse sand and one isolated rounded pebble (Fig. 5). I interpret the fine sands to be lacustrine and the overlying coarse sand and gravel as being fluvial. If this interpretation is correct, the lake elevation in this area is marked by the transition from fine to coarse sand, 1.4 m below the terrace level, ~340.1 m (1,116 ft). Exposures in the vicinity of the road intersection opposite the Mill Pond dam show that the lacustrine sediments are extensively deformed where they are cut by faults and are slumped indicating that the lacustrine and later fluvial sediments in this area were deposited on top of pockets of dead ice. This implies that the lake formed synchronously with the final melting of glacial ice from the valley.

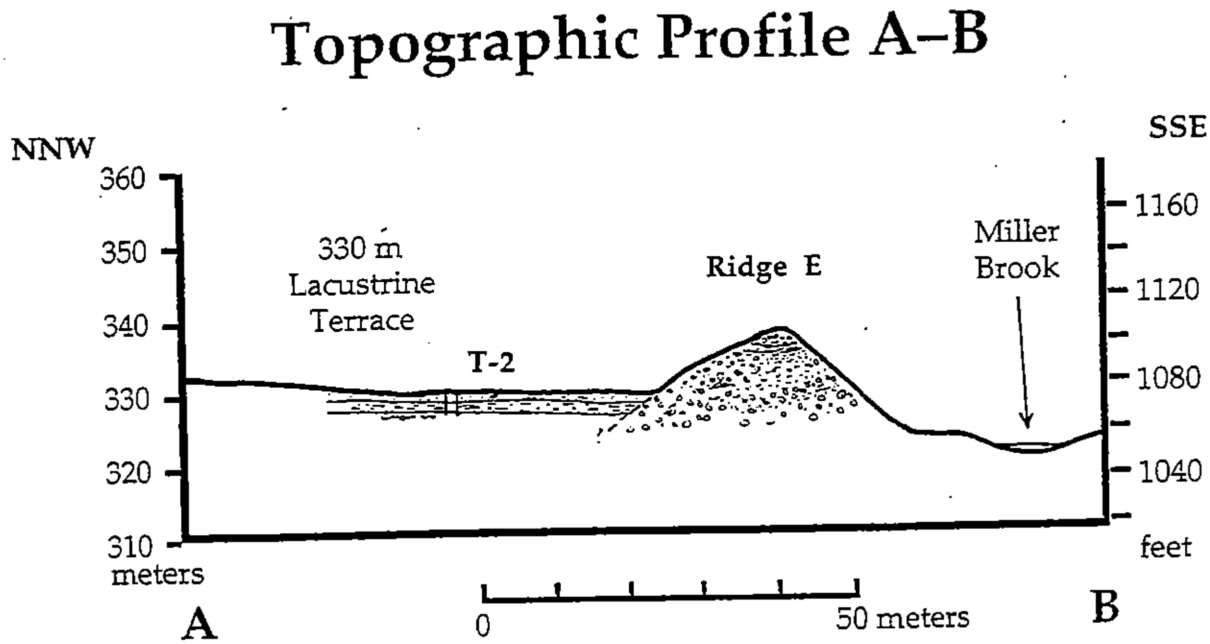
The lower 330 m terrace also appears to record both a lacustrine and fluvial activity. Soil Pit T-2 (along cross-section A–B) was excavated to 1.60 m where it bottomed in coarse rounded gravel and sand. These are overlain by 0.85 m of fine sand and silt which in turn is overlain by 0.75 m of rounded gravel and coarse sand (Figs. 4 and 5). The lake elevation here is interpreted to be at the contact between the fine sand/silt and gravel, or at ~329.2 m (1,080 ft).

The two terraces, corresponding to lake elevations of 340.1 and 329.2 m, may correlate with Lake Winooski, the outlet of which lay 4.0 km south of Williamstown, Vermont at an elevation of 279 m (Larsen, 1987b; see also review of regional glacial lakes later in this field guide). Projecting this threshold elevation N21.5W to the Miller Brook valley with an isostatic gradient of 0.9 m/km gives an elevation of 322.2 m, implying a water depth at the outlet of 7 m for the lower lake level and almost 18 m for the higher lake level, unless the outlet eroded from its current elevation.

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 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, continued SSE ice flow 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 (elev. 103 m, 340 ft), (Fig. 1). 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 years ago (Lin 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 work down-valley 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 imply that that 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.

Wagner's (1970) interpretation that the ridges described here are moraines stemmed from the following observations: (1) The head of Miller Brook occurs in a bowl-shaped valley, steep-sided and flat-bottomed, a shape typical of glacial cirques; (2) The ridge extending down-stream and along the south side of the valley from the Lake



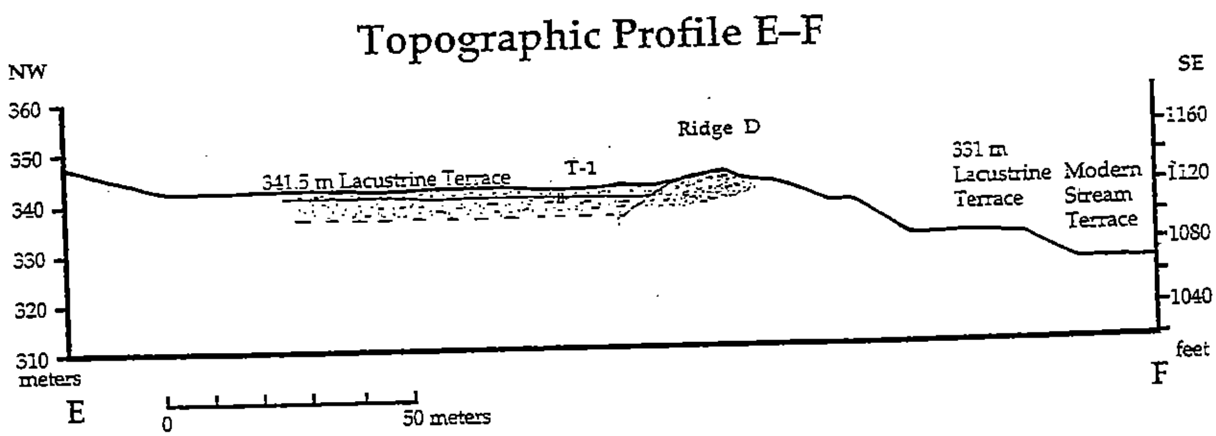


Figure 5: Unexaggerated topographic profiles across terraces and Ridges D and E northeast of the Mill Pond (Fig. 4). **Profile A–B** extends from the 330 m terrace, across **Ridge E**, to Miller Brook. Ridge E is composed of well sorted and bedded coarse sand and gravel. **Profile E–F** extends from the 341.5 m terrace across **Ridge D** to a terrace at 331 m and then down to alluvial terraces adjacent to Miller Brook. Ridge C consists of rounded and bedded gravel and cobbles, similar to Ridge E. Materials exposed in Ridges D and E are interpreted here to be of fluvial origin deposited in an esker tunnel. Soil pits near the border of Ridge D and the 341.5 m terrace show the terrace sands and gravels onlapping the much coarser cobble gravels contained in the Ridge. Soil Pit T-1 in the 341.5 m terrace exposes 1.4 m of medium to coarse sand and gravel that overlie at least 2.2 m of fine sand. Soil pit T-2 in the 330 m terrace exposes 75 cm of coarse sand and gravel overlying 85 cm of grey silt and fine sand, which in turn overlies coarse gravel. The upper sand and gravel in both terraces is interpreted to be fluvial in origin, whereas the underlying fine sand and silt lacustrine deposited in lakes at elevations of 340.1 and 329.2 m respectively.

Mansfield dam turns and crosses the valley in a manner similar to lateral and end moraines produced by alpine glaciers; and (3) A sloping terrace, extending from the end moraine down-valley to a delta was interpreted to be an outwash plain graded from the end moraine down to a pro-glacial lake. I concur with Wagner (1970) that the head of the Miller Brook valley is cirque-like, especially its uppermost reach (Fig. 3), despite arguments by Waitt (Waitt and Davis, 1988) to the contrary. While Loso and others (1997) have determined that the floor of the Miller Brook valley is too low to maintain a valley glacier after the retreat of the Laurentide ice sheet. Given our current understanding of climatic cycling during the Quaternary where gradual cooling over has been followed by rapid warming (e.g. Dansgaard et al., 1993), it is likely that the Miller Brook cirque, and others in New England, formed during the onset of these 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 bedrock landform, the cirque at the valley head, formed before the onset of continental glaciation, whereas the ridges of surficial material were formed as the ice sheet thinned and stagnated in the Miller Brook valley.

Conclusions

(1) Detailed mapping reveals that the ridges of surficial material occurring in the Miller Brook valley can be interpreted as a connected system of ridges lying both along the valley edge and valley center.

(2) Ridges B, C, D, and E all contain clean sand and gravel interpreted here to be of fluvial origin. Materials exposed in shallow holes in Ridges A, F, and G are poorly sorted, yet are dominated by relatively coarse rounded clasts. Based on both the map pattern and the fluvial materials exposed within them, I interpret all of these ridges to be eskers.

(3) The map pattern indicates that Ridges D, E, and F were once continuous and are cut by, and therefore younger than, Ridges C and G. It is also possible that the esker tunnel bifurcated and rejoined and that all of the ridges formed synchronously.

(4) Soil pits in the terrace sands commonly expose highly disrupted bedding indicative of collapse following ice melt-out. Closed depressions within and adjacent to Ridges A and B are also likely to be ice melt-out features—kettles. The eskers, kettles, and collapse structures all suggest sediment deposition in contact with glacial ice, most likely stagnant ice.

(5) Onlapping contact relationships of lake sediments over the materials in the Ridges indicate that a lake occupied the valley soon after ice retreat forming terraces at two different elevations. Lake elevations are interpreted to be at 340.1 and 329.2 m, based on the contact between lacustrine and fluvial sediments and may correlate with Lake Winooski.

(6) If Ridges A and B are indeed moraines (Wagner, 1970), they must post-date the eskers, because the eskers are at elevations very close to that of Ridges A and B, necessitating ice, probably stagnant ice, of at least that thickness. If they post-date the melt out of the Laurentide ice sheet, they must have been produced by either (1) a considerable thickening of the Laurentide ice on the West side of the Green Mountains, allowing an ice tongue to extend through the Nebraska Notch, or (2) the establishment of a small cirque glacier. Both require a considerable cooling of the regional climate which is not supported by paleotemperature estimates during the early Holocene (Loso et al., 1997).

(7) If the "cirque-like basins" at the head of the Miller Brook valley are indeed cirques, they most likely formed during the long periods of cooling prior to the advance of the Laurentide and earlier ice sheets.

PROGLACIAL LAKE AND RIVER HISTORY IN THE WINOOSKI DRAINAGE BASIN

Timothy N. Whalen

Introduction

When the Hudson-Champlain lobe of the Laurentide ice sheet retreated from this mountainous region, proglacial lakes developed between the ice and the valley walls, leaving behind a scattered record of deltas and shorelines. As the lakes withdrew from the valleys, rivers began to incise the lake deposits. The subsequent fluvial activity in the Winooski Drainage Basin removed much of the glacial deposits and lowered the valley floors more than 40 m in the lower reaches. The purposes of this paper are 1) to provide an overview of the history of proglacial lakes occupying the Winooski River Valley and its tributaries, with special attention given to the Huntington Valley, 2) to identify when the transition occurred between lacustrine and fluvial processes in each of these valleys, and 3) to describe the mechanisms responsible for river incision in the Winooski Drainage Basin.

Deglacial History

The deglacial record of the Winooski Drainage Basin is one of northwestward ice retreat, ice lowering, and decreasing proglacial lake levels. Larsen (1972; 1987a & b) has previously described the sequence of proglacial lakes in the Winooski Drainage Basin and his five stage model (Table 1) is used as the basis for this discussion. Larsen specifically addressed the chronologies in the Mad, Dog, and Stevens Branch Valleys, but work in the Huntington River Valley (Wagner, 1972; Bryan, 1995) and in the Little River Valley (Merwin, 1908; Connally, 1972) also demonstrates that the history of proglacial lakes in these valleys can be placed within the context of Larsen's model. The following section focuses on the proglacial lake levels in the Huntington Valley and the reader is encouraged to review Larsen (1972; 1987a & b) for specifics concerning the other valleys in the Winooski Basin and Chapman (1937) for the Champlain Valley lake levels.

Table1: Winooski Drainage Basin Proglacial Lake Levels and Ice Positions

Stage	Winooski River	Huntington River	Little River	Mad River	Dog River	Stevens Branch
I	Ice Retreats to Valley Mouths	Lake Jerusalem to Southern Divide	Ice and Local Lakes	Lake Granville to Granville Gulf Divide	Roxbury Divide	Lake Williamstown to Williamstown Divide
II	Ice Retreats to Jonesville	Lake Huntington to Hollow Brook Divide	Lake Winooski above Divide	Lake Winooski to Warren	Lake Winooski to South Northfield	Lake Winooski to Divide
III	Ice Retreats to Richmond	Lake Huntington to Hollow Brook Divide	Lake Mansfield I to Divide	Lake Mansfield I to Moretown	Fluvial	Fluvial
IV	Ice Retreats to Green Mountain Front	Lake Mansfield II to Hollow Brook Divide	Lake Mansfield II to Stowe	Fluvial	Fluvial	Fluvial
V	Lake Coveville to Waterbury	Lake Coveville to Huntington	Lake Coveville to Moscow	Fluvial	Fluvial	Fluvial

Note: The lake positions are cited as the farthest upvalley location for stage and ice postions are cited as position at end of stage.

Stage I lakes were limited to single north-draining valleys because they were bordered to the north by the retreating ice. These lakes drained over southern spillways, located at different elevations, and into the Champlain (Huntington Valley) and Connecticut (Mad, Dog, and Stevens Branch Valleys) Basins. The Stage I spillway in the Huntington Valley is located at the southern divide at an elevation of 460 m (1510 ft), controlling a previously unnamed lake Stage I lake (Fig. 6). Following Larsen's (1972) criteria for naming Stage I lakes based on the nearest village, the name Lake Jerusalem is suggested. Evidence for Lake Jerusalem includes clay deposits found at 240 m (787 ft) south of Huntington Center (Bryan, 1995). During Stage II, the lakes east of the Huntington Valley coalesced, from southeast to northwest in a time-transgressive manner following the retreating ice, into Lake Winooski and utilized a 279 m (915 ft) spillway south of Williamstown, the lowest Stage I outlet. In the Huntington Valley during Stage II, the ice retreated past the Hollow Brook Valley, allowing Lake Jerusalem to empty into the Champlain Valley and to lower to the Lake Huntington level. Lake Huntington drained across a threshold presently located at 204 m (670 ft) in the Hollow Brook Valley, and its waters entered Lake Coveville in the Champlain Valley at the South Hinesburg delta.

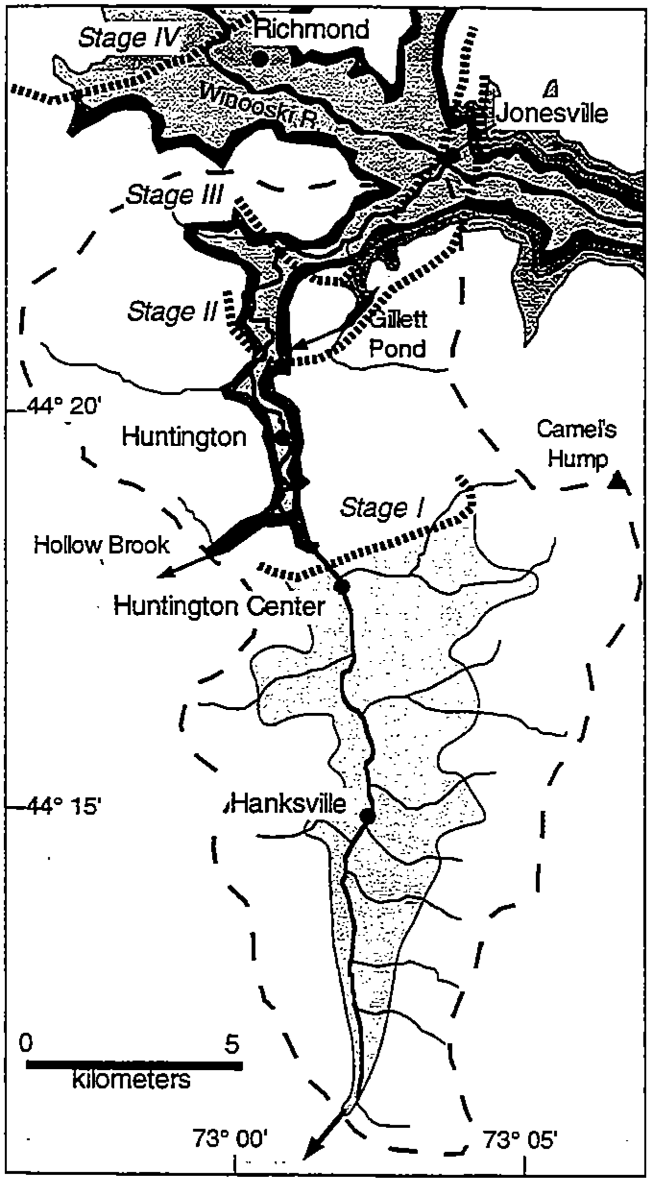


Figure 6: Map of proglacial lake extent and ice positions in the Huntington Valley. The lake levels are shown with the corresponding ice position at the end of each stage.

At the close of Stage II, the ice in the Winooski Valley had retreated to Jonesville, and Lake Huntington had expanded to fill the valley between the Hollow Brook spillway to the southwest and Gillett Pond to the northeast (Fig. 6). However, a tongue of ice remained in the lower Huntington Valley, and when the Gillett Pond spillway (229 m; 750 ft) was uncovered, Lake Winooski drained into the Huntington Valley. During Stage III, the Gillett Pond threshold controlled the level of Lake Mansfield I, and the Hollow Brook threshold continued to control the water level of Lake Huntington (Fig. 7). Stage IV commenced when the mouth of the Huntington Valley became ice free and the waters of Lake Mansfield I were able to drop to the level the Hollow Brook spillway, forming Lake Mansfield II. Once the ice retreated from the Green Mountain Front (Stage V), Lake Mansfield II lowered, Lake Coveville entered the Winooski Valley, and the present configuration of the Winooski River drainage was established.

Fluvial History

The river history in the valleys of the Winooski Basin began when the last proglacial lake shoaled and therefore commenced at different times in each valley. The upper reaches of the Mad River (south of Warren), for example, were clear of lake water during Stage I while the lowest reaches of the Little River remained submerged until the end of Stage V. However, a general valley evolution can be illustrated as in Figure 8, where the initial landscape over which the river flowed was most likely a glaciolacustrine fill terrace. This deposit is generally a coarsening upward sequence of rythmically bedded silts and clays grading into sands (some of which may be deltaic in origin and therefore exhibit foresets) and capped by fluvial deposits. These glaciolucustrine fill terraces have often been mapped as deltas on the basis of their morphology (broad, flat terraces), with little supporting sedimentological evidence (i.e., Wagner, 1972). Precise surveying demonstrates that the elevation of these surfaces decreases downstream in the Huntington River Valley (Whalen, 1997) whereas it should increase (due to glacio-isostatic rebound) if they were all built to a stable lake level. In fact the glaciolacustrine fill terraces are a composite of many depositional environments and record the transition from lacustrine to fluvial deposition.

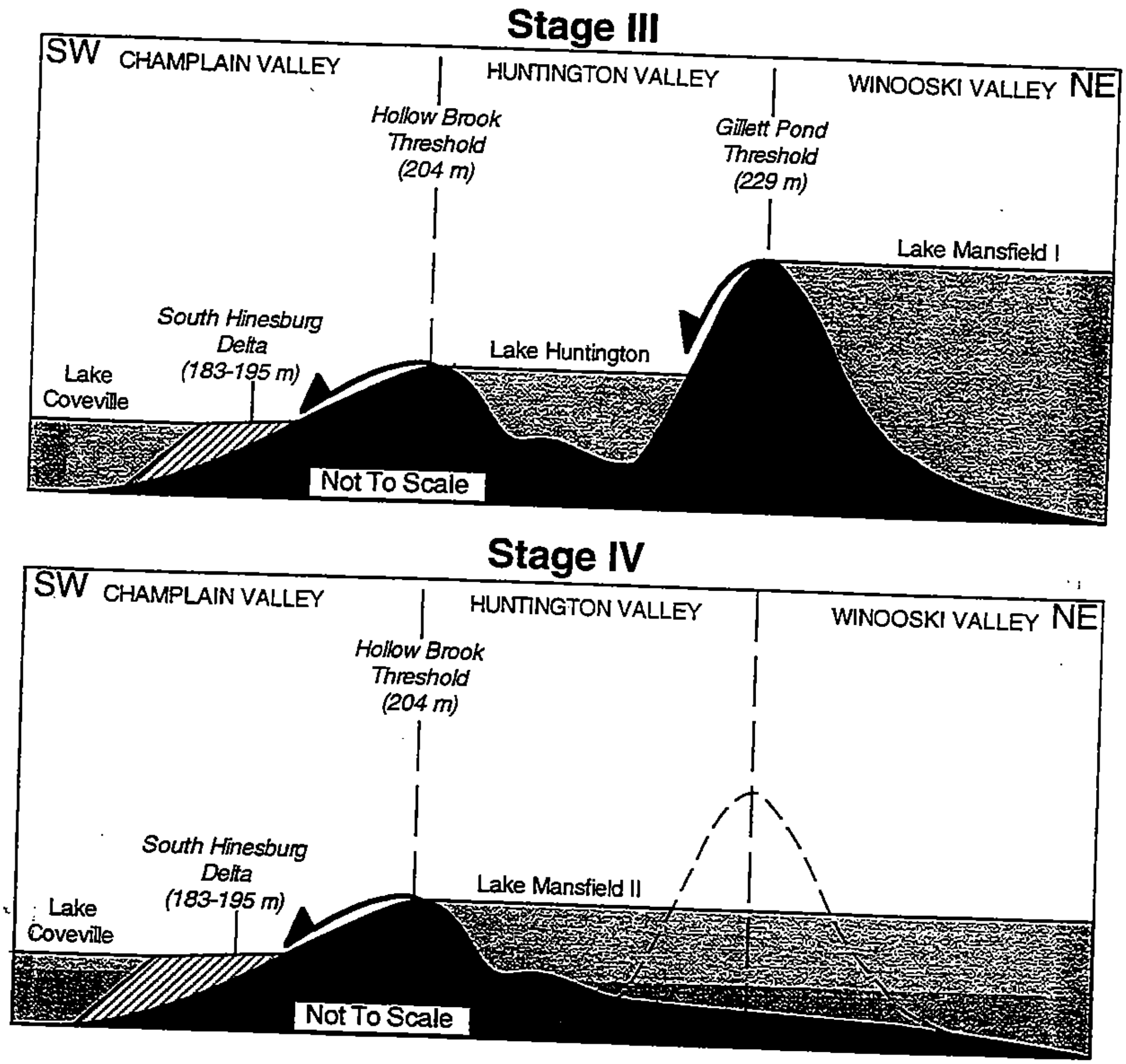


Figure 7: Schematic cross-section through the Huntington Valley during Stages III and IV. During both stages, Hollow Brook was a spillway into the Champlain Valley. During Stage III, when ice blocked the mouth of the Huntington Valley, Lake Mansfield I was forced to drain through the Gillett Pond Spillway. See Figure 6 for ice positions and lake extent.

Subsequent to the valley filling by the glaciolacustrine terrace, dramatic and episodic incision took place. Incision was a response to the proglacial lake base-level drops in the lower reaches of the Winooski Basin. The amount of incision of the glaciolacustrine fill terrace varied along a valley and depended on two factors: 1) the magnitude of base-level change and 2) the distance from the base-level change. Base-level changes continued as first the proglacial lakes, and later the Champlain Sea, in the Champlain Valley lowered to near the present level of Lake Champlain (i.e., Chapman, 1937). Each drop in water level initiated a wave of incision that propagated upstream until a bedrock knickpoint was encountered or until the incision diffused due to a decrease in stream power. Incision resulted in the cutting of the valley fill followed by a period of flood plain development at a lower level; therefore, fill-cut terraces were formed. If a knickpoint lay downstream, the base-level change could not propagate past it and no incision would occur upstream. Base-level changes no longer caused incision after the Upper Marine interval of the Champlain Sea because knickpoints developed along the lower reaches of the Winooski River below Essex.

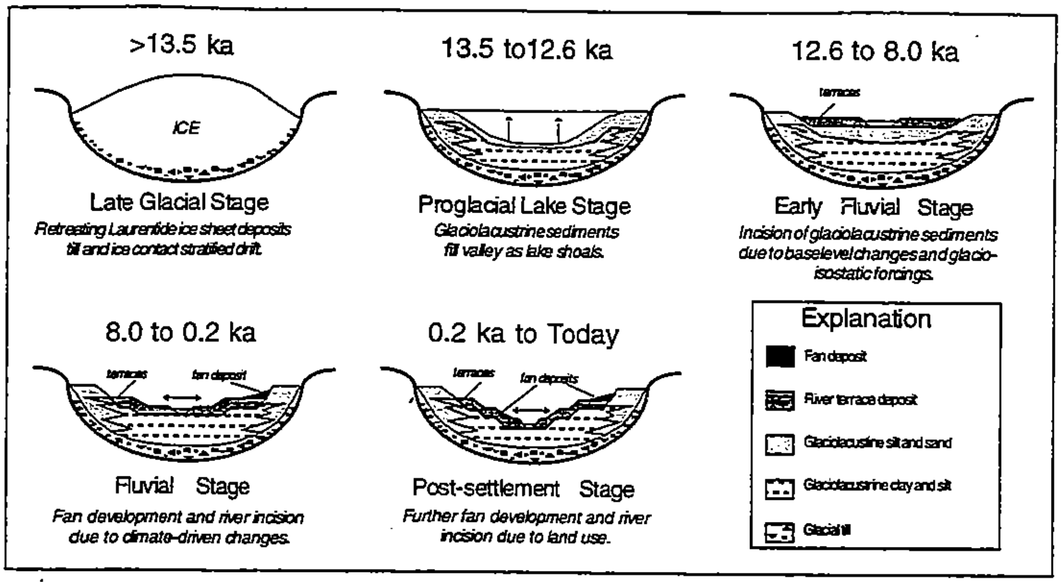


Figure 8: Schematic valley evolution for Winooski Drainage Basin. All ages are in ^{14}C years and are approximate. See text for discussion. Adapted from Church, 1997.

Subsequent to the base-level-driven incision, incision occurred as a response to climate changes and/or internal adjustments of the river system. Widespread climate changes occurred throughout the Holocene and appear to be correlative with periods of fan activity and terrace formation. The relief between terraces decreases after the period assigned to base-level changes which suggests that the mechanism that initiated incision was different. The morphology of the terrace surfaces also changes from a relatively flat feature to terraces dominated by ridges and swales.

Previous Studies of River Terraces in Vermont

One of the only studies of river chronologies in Vermont is that of Brakenridge et al. (1988) who excavated trenches on a point bar of the Missisquoi River, ~60 km north of here, in conjunction with an archeological study. They recovered many datable and identifiable wood fragments from the trenches and they recorded detailed stratigraphy of the terrace deposits. By plotting the channel elevation versus time, their results showed that incision had nearly ceased by 8000 ^{14}C yBP at which point lateral channel migration became the dominant process operating in the river. A period of incision coincides with the time of European settlement in the region (~150 years ago) and presumably the change in river behavior is associated with the clearing of the forests in the watershed for farming and timber.

Brakenridge and others (1988) attributed the decrease in the incision rate at 8000 ^{14}C yBP to be evidence that glacio-isostatic rebound had ended by that time, similar to a time frame proposed by Hutchinson et al. (1981) based on an "undeformed" or horizontal unconformity in the sediments of Lake George that was formed before 7000 ^{14}C yBP. Simple exponential decay models demonstrate that the rate of rebound generally would have decreased to immeasurable amounts during the early Holocene (Whalen, 1997), but episodic incision followed by terrace development continued during the Holocene in the Winooski Drainage Basin. Therefore, the rate of rebound can not in itself control incision, contrary to the suggestion of Brakenridge and others (1988), and other mechanisms must be considered.

Bedrock knickpoints are found in every river, including the Missisquoi below Brakenridge and others' site. Knickpoints stop incision initiated by base-level changes downstream and form local base levels by controlling the river elevation upstream. Even with continued glacio-isostatic uplift during the Holocene, the rate of incision would be dependent on the rate of knickpoint migration upstream. The migration of knickpoints upstream tend to push the fluvial system towards disequilibrium, and periodic internal adjustments, possibly initiated by large floods or widespread climate changes, are evident by the continued formation of terraces during the Holocene. The terrace chronologies in the Winooski Basin reflect a non-synchronous decrease in incision rate among valleys, suggesting that different knickpoints influenced individual valleys at different times.

Moultroup and Aldrich Farms

The flight of terraces at the Moultroup and Aldrich farms (Fig. 9) 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 and represent 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 glaciolacustrine fill terrace (T8). The second stage is dramatic incision followed by deposition of fill-cut terraces as a response to base level changes (T7-T6). The final stage is periodic incision related to internal adjustments of the fluvial system triggered by environmental changes (T5-T1).

Alluvial fans present on T5, T2, and T1 in this section of the Huntington Valley were dated by Church (1997) and Zehfuss (1996). The basal dates define the time of initial fan formation, but also provide lower limits to the time of terrace formation. The clustering of dates around distinct periods for fans built on a given terrace and the consistent relationship of younger fans built on lower terraces suggest that terrace formation is at least in part responsible for the timing of fan formation. In addition, limiting ages of the terrace deposits are reported by Whalen (1997) based on terraces correlated to dated base levels and direct dating of the terrace deposits (Tables 2 and 3).

Table 2: Huntington River Terrace and Basal Alluvial Fan Radiocarbon Ages

Terrace	Laboratory #	Material	Depth (m)	^{14}C Age	Calibrated Date (yr B.P.)	1 Sigma Range (yr B.P.)	2 Sigma Range (yr B.P.)
T1*	GX-21329	Wood	4.00	< 100	62-0	128-0	238-0
T2†	CAMS# 30358	Wood	3.60	1900±50	1830	1879-1751	1939-1710
T2*	CAMS# 22994	Wood	4.00	2500±60	2708-2402	2735-2363	2746-2352
T5†	CAMS# 30353	Charcoal	0.75	7790±60	8424	8541-8407	8565-8367
T5**	GX-20276	Wood	4.00	7835±105	8555	8713-8430	8981-8375
T5**	CAMS# 20963	Wood	1.30	8060±60	8981	8993-8764	9194-8662
T5**	CAMS# 20901	Wood	2.50	8530±100	9486	9531-9436	9819-9275
T6†	CAMS# 30347	Charcoal	0.49	8120±60	8991	9189-8980	9240-8736
T6†	CAMS# 30348	Charcoal	0.46	8230±60	9210	9362-900	9380-8988

Note: * Alluvial fan dated by Zehfuss, 1996. ** Alluvial fan dated by Bierman et al., 1997 and Church, 1997.

† Terrace dated by Whalen, 1997. Calibration using CALIB 3.0.3A (Stuvier and Reimer, 1993).

Two pieces of charcoal from the terrace deposits of T6 yielded dates of between 8100-8300 ^{14}C yBP and provide a minimum age for the formation of T6. However, the age of T6 is believed to be much older based on 1) the correlation of it to the Upper Marine of the Champlain Sea, 2)the ages of the fans built on T5, and 3) the age of the T5 deposits. A basal date obtained from wood below the Audubon fan on T5 is 8530 ± 100 ^{14}C yBP (CAMS# 20901) and another basal date from wood found 4.0 m below the Moultroup fan surface across the valley on T5 is 7835 ± 105 ^{14}C yBP (GX-20276). These dates agree well with the date of 7790 ± 60 ^{14}C yBP (CAMS# 30353) from a piece of charcoal recovered from 0.75 m below the T5 surface, suggesting that the T5 was formed by 8500 ^{14}C yBP and was active until at least 7800 ^{14}C yBP. Therefore, the charcoal recovered from T6 appears to been emplaced after the deposition of T6.

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

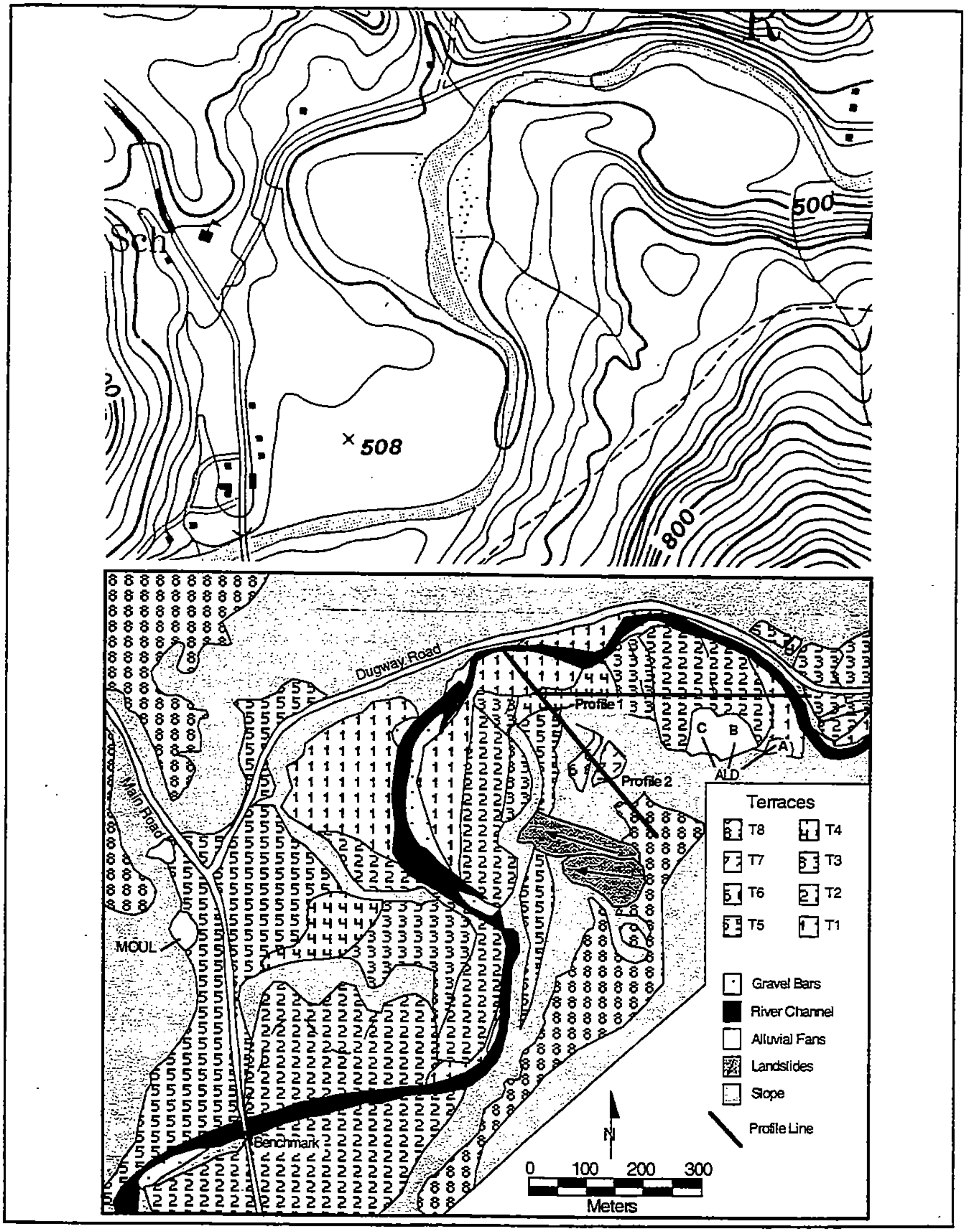


Figure 9: Topographic base map and same-scale map of geomorphic features at Stop 1. See Figures 13 and 15 for detail maps of fans and Figure 10 for profiles of terraces.

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Table 3: Summary of Terrace Characteristics at Moultrop Farm

Terrace	Type	River System	Elevation (m)	Incision (m)	Age (^{14}C yBP)
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	7.8-6.0(?)
T3	Fill-cut Terrace	Meandering	158.90	3.60	6.0(?)-2.5
T2	Fill-cut Terrace	Meandering	155.47	3.43	2.5-0.2
T1	Fill-cut Terrace	Meandering	151.25	4.22	0.2-Today

Note: Ages based on correlations to dated baselevels in Champlain Valley and dated alluvial fan and terrace deposits (Table 2; Whalen, 1997).

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 (Fig. 10). 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 3). The obvious differences in relief between the upper (T8-T5) and the lower (T5-T1) terraces corresponds to the change in incision mechanisms and river behavior that is also visible in the terrace stratigraphy.

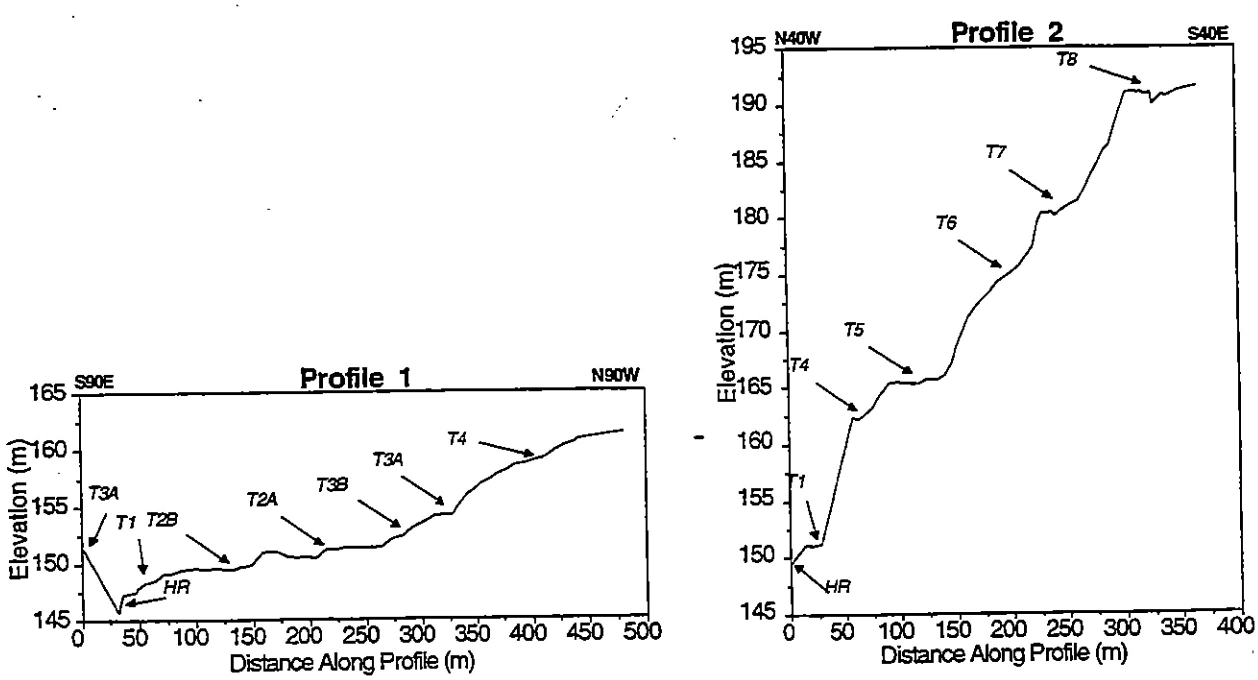


Figure 10: Cross-valley profiles of terraces at the Moultrop and Aldrich farms. Location of profiles is shown on Figure 9. Vertical exaggeration is 10X.

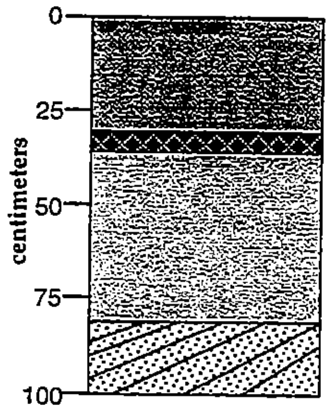
Backhoe trenches and shovel pits, opened on nearly every terrace here, revealed three different terrace stratigraphies (Fig. 11). Common to all terraces is the presence of a 25 cm plow zone capping the fluvial deposits. The plow zone represents the historical land use at this location as tilled fields before the present use as pasture land began about 35-40 years ago (pers. comm., Henry Moultroup). The upper meter of the glaciolacustrine fill terrace, T8, is composed of cross-bedded sand overlain by silty fine sand that suggests deposition by a meandering fluvial system. The stratigraphy of the pits and trenches from T7 and T6 (represented by the T6 log in Figure W5) 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 terraces deposits from T5 to T1 all show a similar stratigraphy, implying that the fluvial system behaved similarly to today's meandering river since the time of T5. The basal unit of imbricated gravels, deposited in the channel, is overlain by laterally accreted sand which is then capped by overbank deposits of silty fine sand.

The changes in terrace stratigraphy from T8 to T1 demonstrate that the Huntington River has evolved through time. T8 developed at the edge of the shoaling Lake Coveville and therefore probably represents a low gradient river, evident by the abundant fine-grained sediments and meandering-type deposits. However, in upvalley locations the glaciolacustrine terrace is capped with gravel-rich deposits that point towards deposition by a braided river; therefore the proximity to the lake appears to control the type of deposit. The fact that T7 and T6, composed of braided river deposits, formed farther from their respective base levels than T8 is consistent with this relationship. All three terraces developed at a time when incision due to base level changes and hillslope instability due to lack of vegetation would have delivered abundant sediment to the river. Initially, the postglacial landscape was out of equilibrium with prevailing conditions. Vegetation was sparse and due to the instability of the hillslopes, abundant sediment was transported to the river. As vegetation spread across the landscape and the hillslopes stabilized, the sediment supply decreased and equilibrated with the dominant processes, and the gradient of the river decreased as rebound lifted the mouth relative to the river. In addition, the decrease in incision after the initial period of rapid base level changes means that less sediment was eroded upvalley. Together, the changes in the type and amount of sediment supplied to the river led to the development of the fluvial system represented by T5-T1.

T5-T1 formed after the base level changes and glacio-isostatic rebound were substantially complete. The relief between the terraces is less than 5 m at this location, and sublevels are prevalent. Upstream of the Moultroup farm, these sublevels are indistinguishable; so the forcing which is responsible for them was diminished upstream. The proposed mechanism for the formation of these terraces is periodic incision as a response to knickpoint migration. Two kilometers downstream begins the Huntington Gorge. The last terrace to grade across the uppermost section of the gorge is T4. Therefore, the evolution of the gorge through the migration of the knickpoint upstream is a prime mechanism for causing the incision below T5.

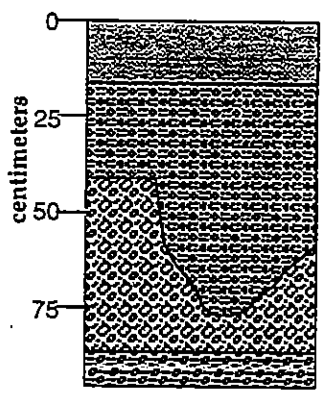
The terraces at the Moultroup and Aldrich farms provide one of the best examples of the different forms of fluvial terraces found in the Winooski basin. The terrace chronologies discussed here are consistent with those studied in the Little and Mad River valleys and may be applicable to other terraced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

T8



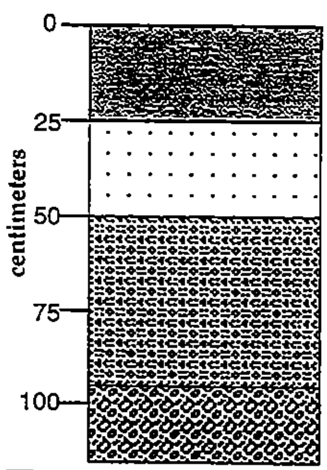
Unit	Sediment	Color	Interpretation
I	Fine Sand with Pebbles	Yellowish Brown 10YR 5/6	Plow Zone
II	Silty Fine Sand	Light Grey 10YR 7/1	Leached Overbank Deposits
III	Silty Fine Sand	Olive Yellow 2.5Y 6/6	Overbank Deposits
IV	Cross Bedded Coarse Sand	Olive Yellow 2.5Y 6/6	Point Bar Deposits

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

T5



Unit	Sediment	Color	Interpretation
I	Fine Silty Sand with Pebbles	Light Olive Brown 2.5Y 5/4	Plow Zone
II	Silty Fine Sand	Olive 5Y 5/3	Overbank Deposits
III	Coarse Sand and Pebbles	Olive Grey 5Y 5/2	Point Bar Deposits
IV	Imbricated Gravels with Coarse Sand Matrix	N/A	Channel Deposits

Figure 11: Examples of sedimentological differences in the terrace deposits. These three shovels pits along Profile 1, illustrate the changes in river environment through time. See text for discussion.

ALLUVIAL FAN HISTORY, HUNTINGTON, VERMONT

Paul Zehfuss and Paul. R. Bierman

Introduction

Alluvial fans are seldom studied landforms in the Northeast. We have investigated more than twenty alluvial fans in the Huntington River Basin (Bierman et al., 1997; Church and Bierman, 1994 and 1995; Zehfuss and Bierman, 1995), clustered in the northern part of the river valley (Fig. 12). All of the fans are small (<2500 m^2), and have been deposited where Huntington River terraces provide a platform for fan sedimentation. Three sites along the Huntington River have received the most attention and have been used to infer the regional history of hillslope stability and climate change during Holocene. These three sites; Audubon, Moultroup, and Aldrich, include 5 of the 22 alluvial fans identified in the Huntington River Basin, and have supplied information about fan aggradation, and subsequently hillslope erosion, spanning the last 8530 ^{14}C years. This trip will visit the Moultroup and Aldrich fans.

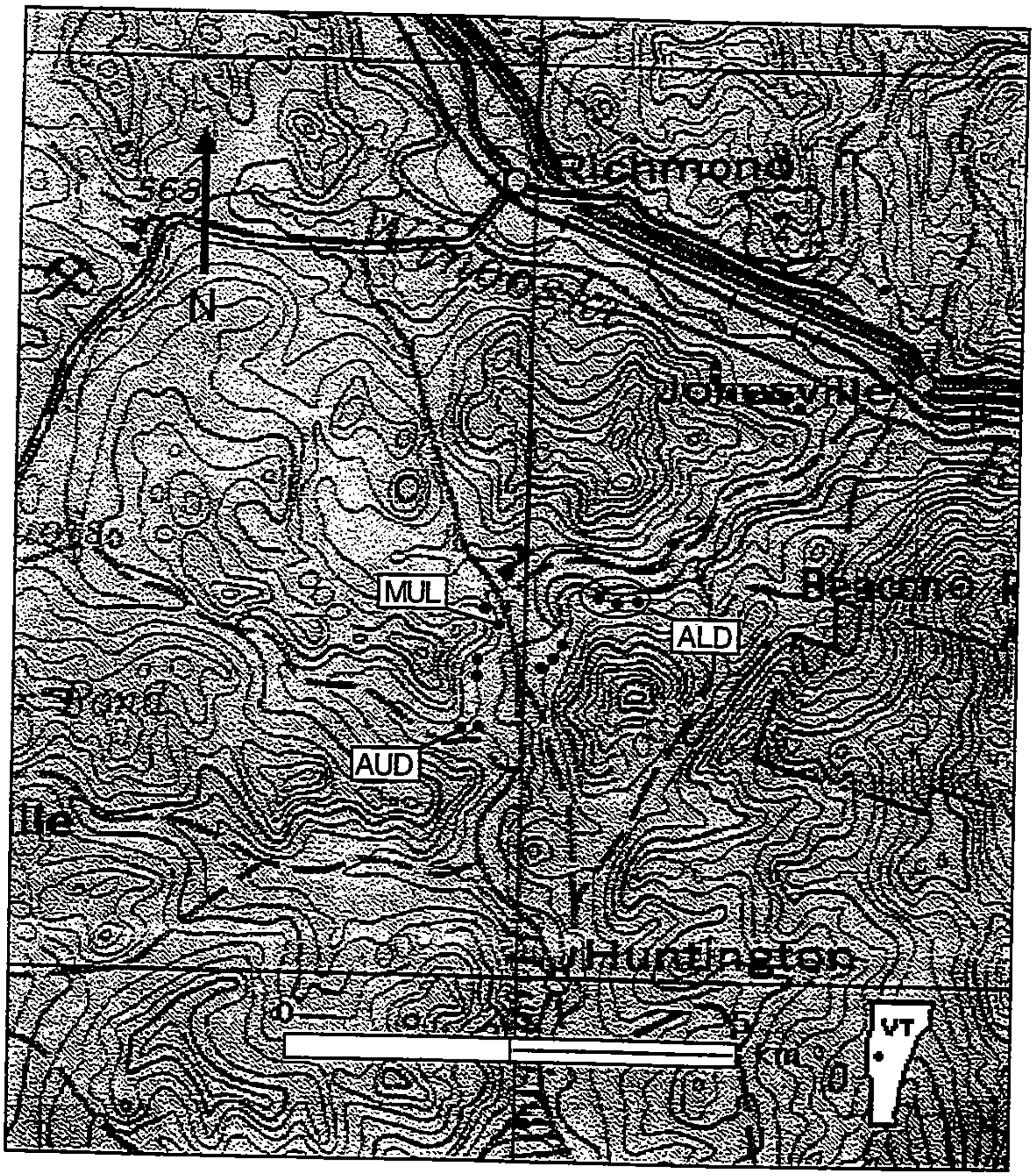


Figure 12: Topographic map showing the location of 22 alluvial fans in the Huntington River Basin. MUL – Moultroup Fan; AUD – Audubon Fan; ALD – Aldrich Fans.

There are similarities between the five fans that we have trenched. Each fan contains a buried soil profile, generally within the upper meter. Radiocarbon dates confirm that the sediment above the organic horizons, often amounting to more than a meter, was deposited coincident with historic destabilization of hillslopes. The soil allows us to identify the period of deposition following the arrival of settlers, who rapidly cleared the hillslopes of trees. A plow zone is usually present above the buried soil, representing the period of 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 methods for obtaining ages of sediment and estimating rates of deposition.

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 mountains, or on slopes created by the edges of higher Huntington River terraces. Very little deposition appears to be occurring on the fans today, although thin (<1 mm) beds of fine silt have been deposited by ephemeral stream activity during severe storms and winter thaws within the past few years.

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, and between 7835 and 7360 ^{14}C yBP), two in the late Holocene (2500 and 1900 ^{14}C yBP), and one fan aggraded over 4 m during historic time (<100 ^{14}C yBP). During the field trip, we will view the fans from which we have obtained aggradation histories, as well as to introduce the types of fans common to this region of the Northeast.

Moultroup Fan

The Moultroup fan is seen on the west side of the Richmond–Huntington road, just south of the right-hand turnoff onto Dugway Road (Fig. 12). There is a second, larger fan to the north, with a house built on its surface. The Moultroup fan was trenched extensively by Church and Bierman in the summer of 1995 (Fig. 13). The fan radius is approximately 65 m, with an apex height of 5 m. A band of tall grass in the center of the field marks the approximate toe of the fan, an area of groundwater discharge. The Moultroup fan has accumulated sediment derived from deposits of glacial till and glaciolacustrine deposits that cover the adjacent bedrock highlands.

Trenching of the Moultroup fan revealed 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. 14). Aggradation rates were rapid in the early Holocene when the fan was beginning to form, between 7835 to 7360 ^{14}C yBP. Sedimentation slowed considerably in the mid-Holocene (7360 to 100 ^{14}C yBP). 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 date (<100 ^{14}C yBP) just above the buried soil indicates that the overlying sediment postdates European settlement and is likely related to land clearance and agricultural practices.

The Aldrich Fans

The Aldrich fans are found on the south side of the Huntington River along Dugway Road, about 1/2 mile east of the Moultroup Fan (Fig. 12). From the river looking south, two fans are readily apparent in the center of the far side of the Aldrich field (Fans B and C, Fig. 15). The fans extend almost 80 meters into the field, and coalesce with one another in the mid-fan portions. Farther east, the smaller Fan A is less obvious (Fig. 15). The Aldrich fans have been deposited on a terrace of the Huntington River that is characterized by uneven terrain that varies in relief by about one meter. The Huntington River marks the northern boundary of the Aldrich farm, where the river has incised nearly 2 m below the surface elevation of the field. Swales between higher deposits of sediment are often filled with puddles following rainstorms and snow melting events.

A steep slope, formed by a higher Huntington River terrace, is the source of fan debris. Drainage basins cut into this higher terrace contain ephemeral streams leading to the fans below and are filled with ice and flowing water in the winter and spring months. Groups of very small stream beds on the terrace top converge at the head of the drainage basins. During winter, thin layers of fine, tan silt are suspended on top of ice which can be seen coating the surfaces of the fans. Clasts ranging in size from very large cobbles to small pebbles line the ephemeral stream beds leading to the fans. Deposits of sand and silt have been exposed along the steep sides of the drainage basins, and in some places have plunged into the stream beds as small land slides (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.

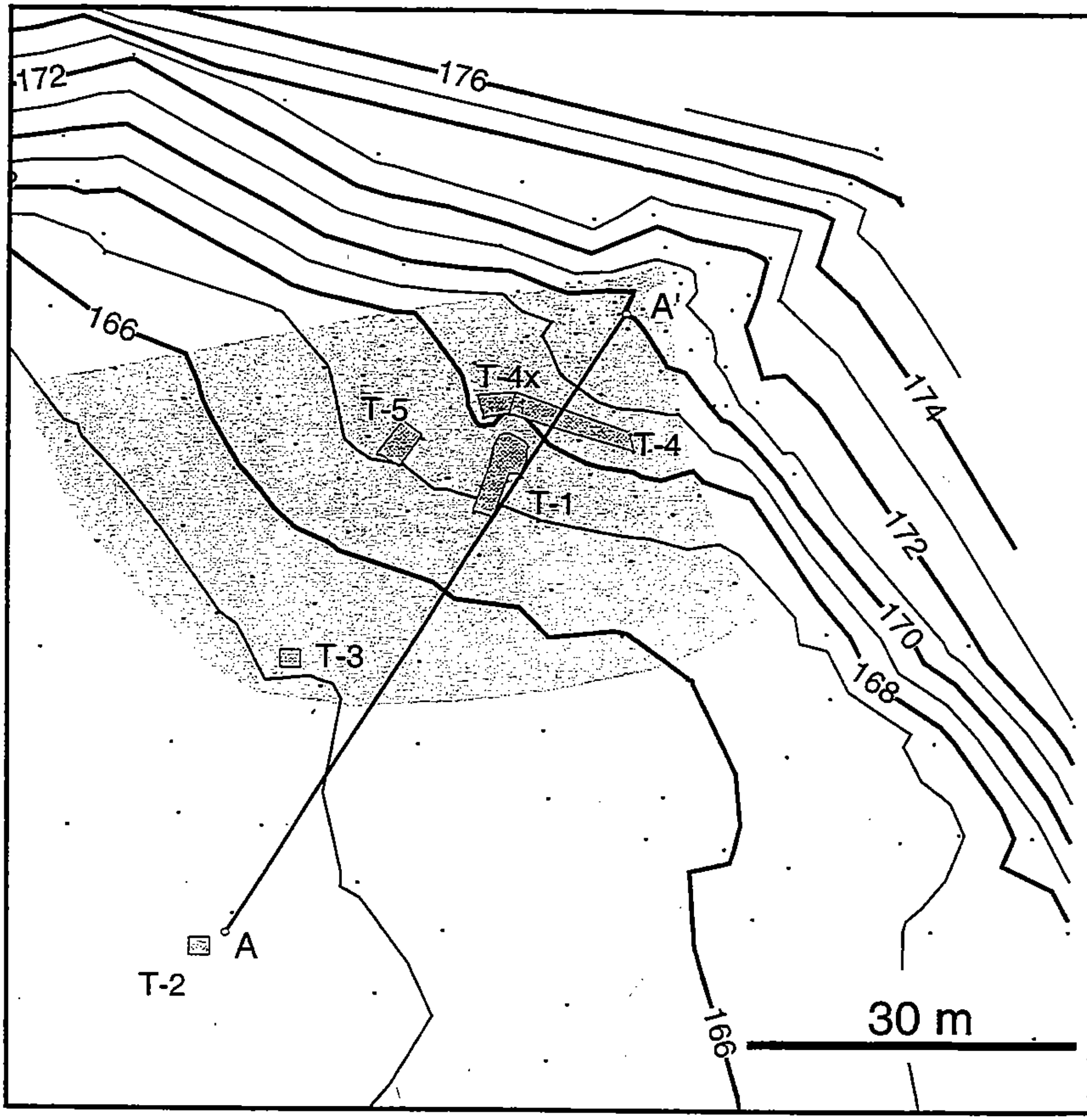


Figure 13: Topographic map of the Moultroup Fan (shaded) showing total station survey points, location of trenches (T-1 to T-5), and line of cross-section (A to A'). North is to the bottom of the map. Contours are in meters above sea level.

Fan A

Fan A is the closest to the Huntington River of the group of three Aldrich fans (Fig. 15) and is predominately composed of sediment deposited after European settlement. A large chunk of wood recovered from four meters below the surface of Fan A was radiocarbon dated at < 100 ^{14}C yBP. Thus, the four meters of sediment above the wood is interpreted to be post-settlement fan deposition. The trench was positioned along the fan surface where there is approximately 5 meters of total fan deposition. This implies that nearly 80% of the vertical aggradation at this location on fan A is composed of post-settlement fan deposition.

Fans B and C

Five trenches were excavated into fans B and C and logged (Fig. 15). Radiocarbon dates provide basal ages for the fans (Fan B, 1900 ^{14}C yBP; Fan C, 2500 ^{14}C yBP), as well as time markers within the stratigraphy exposed by the fan trenches. Figure 16 is a trench log from Fan B. A significant period of relative fan surface stability is marked

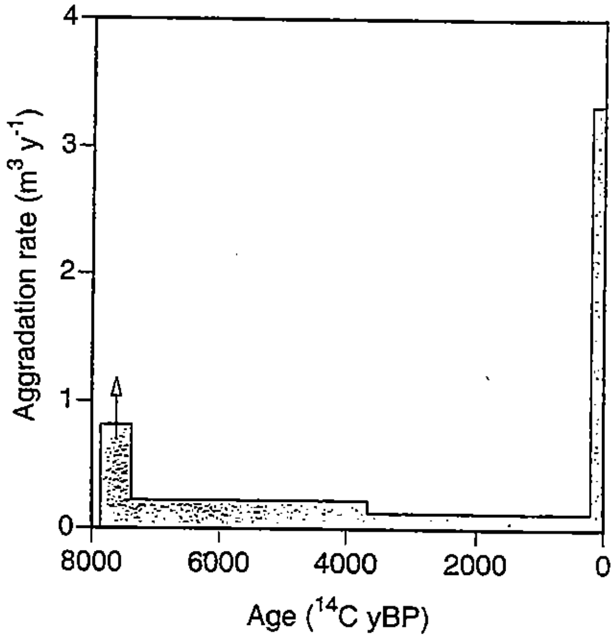


Figure 14: Estimated aggradation rates in ^{14}C years to the Moultroup Fan (assuming that fan geometry is that of a right circular cone) demonstrate response of fan during historic period (180 years, based on settlement history).

by the paleosol A horizon identified in most Aldrich trenches. The buried soil represents a period when vegetation became well established on the fan surface, and also indicates that hill slopes were supplying little sediment to the fan. Radiocarbon dates of 230 and 310 ^{14}C yBP obtained from wood and charcoal in the paleosol A horizon (Fig. 16) allow the soil horizon to be correlated between trenches (Fig. 17) and further demonstrate the fan sedimentation during the time of European settlement. We suggest that the extensive deforestation that occurred during settlement in the late eighteenth century destabilized the hillslopes. Without stability provided by root systems, the soils and sediments were more easily eroded from drainage basins, increasing sedimentation on the fan surfaces.

Following deforestation, sediment continued to be deposited at a relatively rapid rate as compared to depositional rates previous to settlement. As the farming industry began to develop in Vermont, the Aldrich field was utilized for agricultural purposes. Verbal communication with residents of the Huntington and Richmond regions has verified that the field was plowed repeatedly previous to 1960. A cross-section of Fan B (Fig. 17) 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 may have existed throughout the entire Aldrich field at one point, but was combined, where it was not buried by alluvial fan sediments, with the cumulative plow layer during farming. Following the cessation of row-crop farming on the Aldrich meadow and its subsequent and continued use for hay and pasture, the cumulative plow layer has been buried by continued fan deposition (Fig. 17). The recovery of forest growth on the hillside has returned some, but not complete, stability to slopes in the drainage basin.

Conclusions

Our investigation of alluvial fans in the Huntington River Basin has shown that fans provide an accurate, datable record of post-glacial hillslope stability. The sensitivity of fans to their environment, climatic conditions and vegetation, make them useful when quantifying the landscape response to shifts in environmental conditions. When assessing the impact of human-induced landscape modifications, fans are sobering examples of land-use impacts, and remind us of the need for prudent land management.

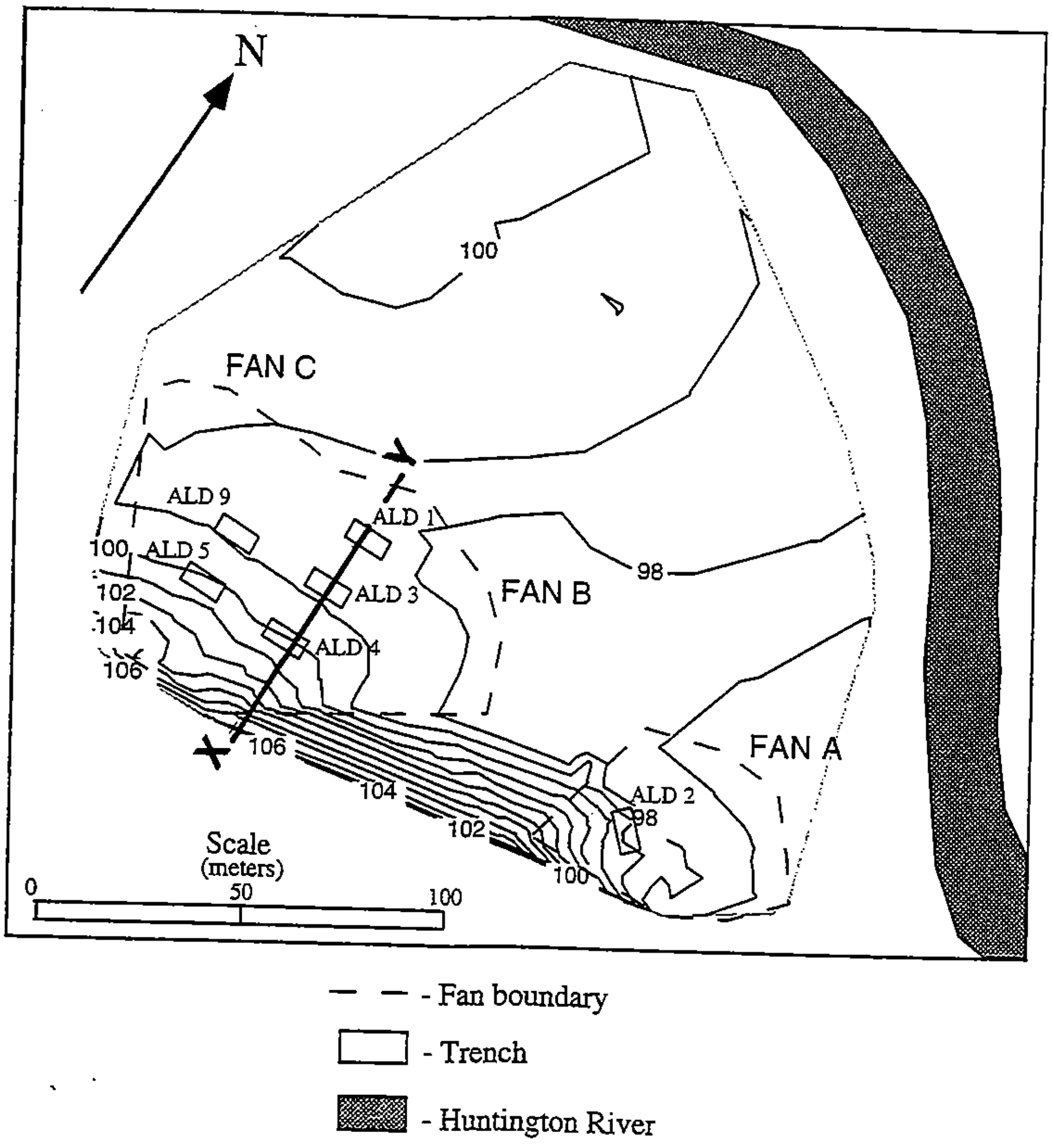


Figure 15: Contour map of the Aldrich field showing the location of 6 backhoe trenches (e.g. "ALD 3"). Alluvial Fans A, B, and C are outlined by dashed lines. Cross-section X–Y through Trenches ALD 1, ALD 3, and ALD 4 (Fig. 17) is shown with heavy line. Contours are in meters above an arbitrary datum.

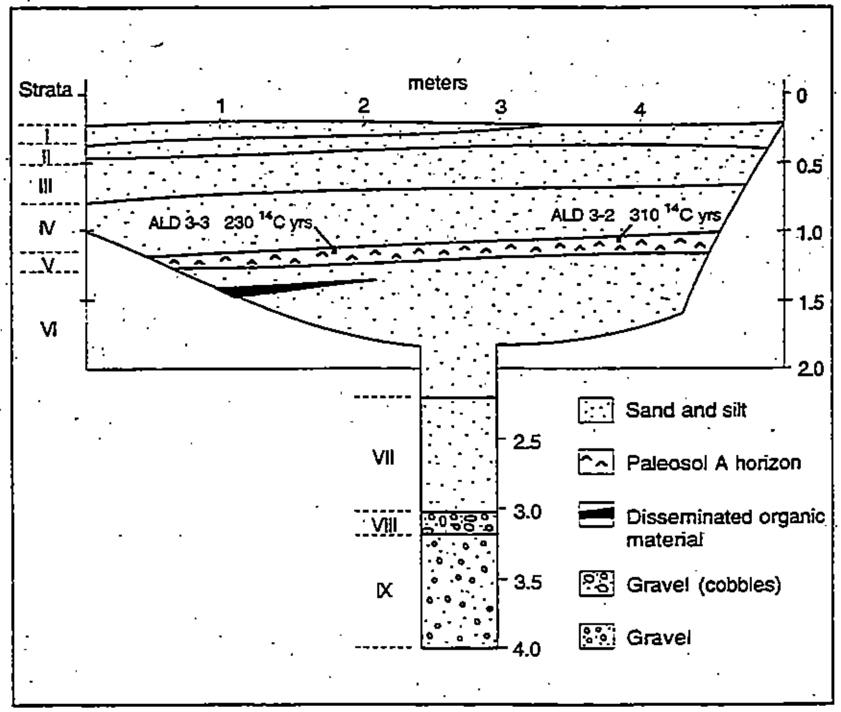


Figure 16: Log of Trench ALD-3 in Fan B (Fig. 15) showing river gravels (Units VIII, IX) underlying fine-grain overbank (Unit VII) and post- and prehistoric alluvial fan deposits (Units I through VI) along with radiocarbon ages on single pieces of charcoal from the paleosol horizon. View is to the south. Vertical scale is in meters and there is no vertical exaggeration.

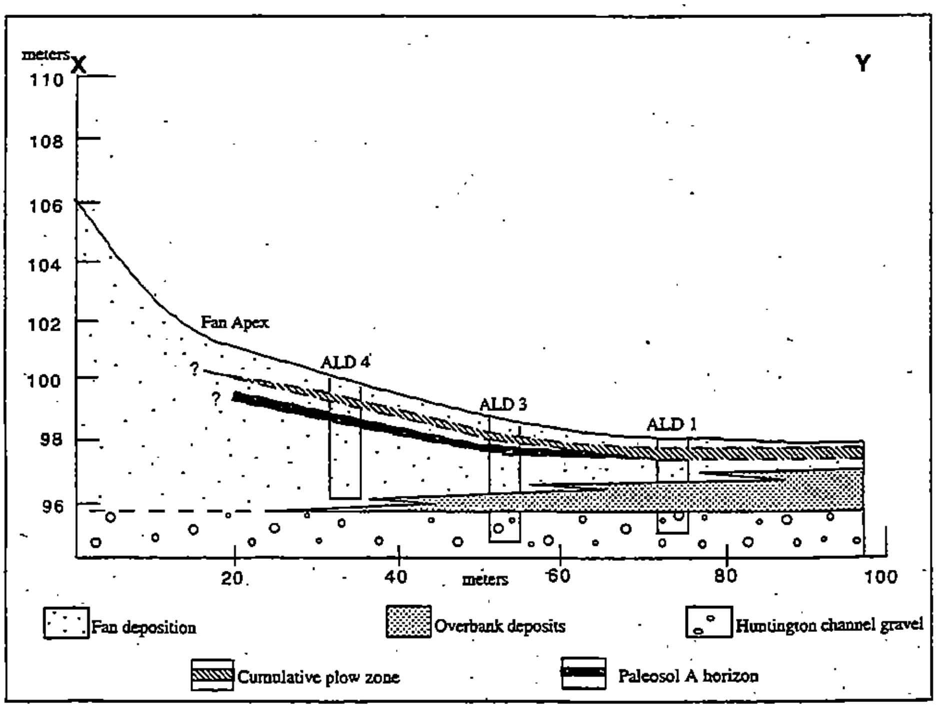


Figure 17: Cross-section of Aldrich Fan C, looking West (Fig. 15), incorporating information from Trenches ALD 1, ALD 3, and ALD 4. Cross-section is drawn with 3.7X vertical exaggeration and elevations are relative to an arbitrary datum.

ACKNOWLEDGMENTS

We are deeply indebted to the willing access granted us by the Moultroup and Aldrich families of Huntington and the Lake Mansfield Trout Club for allowing not only us, but many UVM students to traverse and poke at the fascinating landforms on their properties. We also thank Michael Loso and Howard Schwartz for allowing us to use their detailed map of portions of the Miller Brook valley as a basis for some of the work presented here. Thanks are also extended to Lori Barg for assisting with surveying in the Mill Pond area.

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ROAD LOG

Pico Ski Area to Richmond Exit I-89

Mileage: Cummulative from Pico	Mileage: Cummulative from South Hinesburg	Mileage: Since last turn	
0.0	-	0.0	Start mileage when leaving Pico parking lot, Left on US 4 west
8.8	-	8.8	Right on US 7 north in Rutland
37.3	-	28.5	Right on VT 116 north/ VT 125 east
37.8	-	0.5	Left turn off of VT 125 east, follow VT 116 north straight after stop sign
49.6	-	11.8	Right at stop sign onto VT 116 north, VT 17 west
64.2	0.0	14.6	Right onto Hollow Brook Road, Alternative start at Blaises Country Store in South Hinesburg
67.4	3.2	3.2	Hollow Brook Divide
69.4	5.2	2.0	Left onto Main Raod at stop sign
73.0	8.8	3.6	**Moultrop Farm, STOP 1** (Henry Moultroup: 434-2279)
-			Walk 0.6 miles north from the Moultrop Farmhouse, down Dugway Road to crossing.
73.2	9.0	0.2	Dugway Road
73.4	9.2	0.2	Top of T8
76.7	12.5	3.3	Left before round church, following (To US 2 sign)
77.3	13.1	0.6	Left at light onto US 2 west
78.8	14.6	1.5	I-89 on ramp in Richmond

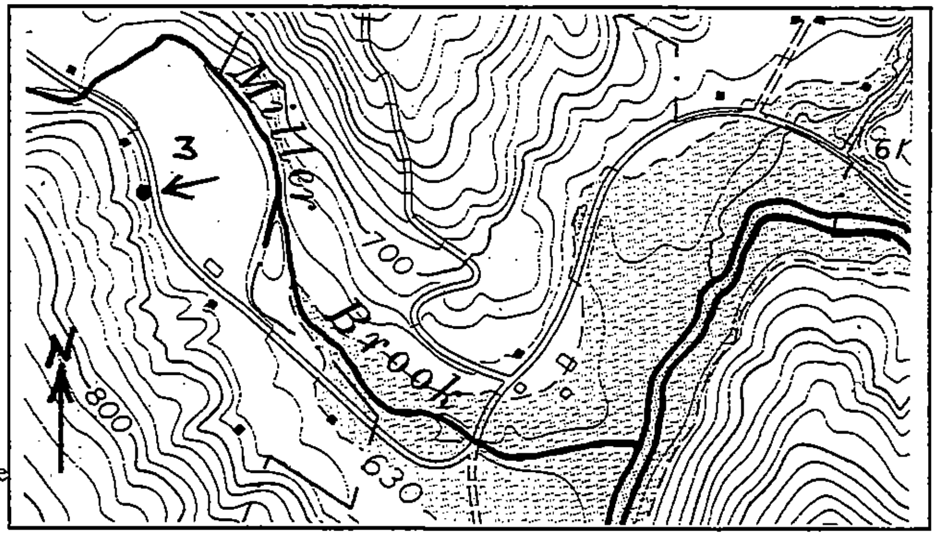
Richmond Exit I-89 to Miller Brook (Stop 3)

Mileage
0.0 **Richmond Exit, I-89:** Travel east along I-89 and exit at Waterbury.

13.7 **Intersection of I-89 and Route 100:** Take Route 100 north 7.1 miles until reaching the turnoff (left or west) for Moscow. The turnoff is marked by a small green sign and arrow.

20.8 **Intersection of Route 100 and Town Road to Moscow:** Follow this road west through the village of Moscow. The road follows the Little River valley west and then south.

22.8 **Bridge over Miller Brook:** Follow the paved road around a sharp right turn immediately after crossing the bridge. A dirt road continues straight along the river valley and eventually dead ends at the Waterbury Reservoir. The road now follows Miller Brook. Continue up the road 0.4 miles. An active alluvial fan is forming on the meadow along the left-hand side of the road. Above map is an enlarged portion of the Stowe 7.5' Quadrangle (1:12,000, North up) showing the location of Stop 3. Miller Brook and the Little River have been highlighted for clarity.



23.2 **STOP 3: ACTIVE ALLUVIAL FAN**
Request permission to visit this site from the occupants of the farmhouse below the fan. One can usually park on the meadow or along the side of the road.
Continue up the Miller Brook Road.

24.5 **Road to gravel pit takes off to left.** Continue straight ahead on the Miller Brook road which turns to dirt shortly after the intersection.

25.0 **Gravel Pit:** Time permitting we will visit this pit on our return down the valley. Continue up the road. Use caution as the road winds and is extensively used by cyclists and pedestrians.

26.4 **Intersection with road to right opposite dam on Miller Brook:** Turn right and follow this road (Old County Road) approximately 0.1 mile to pull out on left that leads to a parking area along the edge of a small meadow.

26.5 **STOP 4: MILLER BROOK LANDFORMS**
Most of the area traversed on this stop belongs to the Lake Mansfield Trout Club. Please request permission before visiting the Trout Club property (253-7565).

END OF TRIP