TRIP A-2:

SLOPE STABILITY AND LATE PLEISTOCENE/HOLOCENE HISTORY, NORTHWESTERN VERMONT

Paul R. Bierman, Stephen F. Wright, and Kyle Nichols, Department of Geology,
University of Vermont, Burlington, VT 05405

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 (Figure 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. Farther east, the Connecticut River flows south to the Atlantic Ocean. 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 are covered primarily with varying thickness of till. Where bedrock does crop out, it is primarily metamorphic. 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 ice sheets throughout the Quaternary. The latest advance probably overran the state sometime before 27,000 ^{14}C y BP (Fullerton, 1986), at its maximum burying northern Vermont under several kilometers of ice. Deglaciation appears to have occurred by thinning, separation over the mountains, step-wise stagnation zone retreat, and eventual larger-scale stagnation and melting 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 northeast, are 800 ^{14}C years younger (Li, 1996), consistent with a thinning ice sheet.

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, ice sheets accentuated the pre-existing landscape, by removing most of the weathered rock, steepening slopes, and reshaping 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 Saint Lawrence River. When ice filled all or part of the Champlain Lowland (both during glacial advance and retreat), north-flowing drainage was blocked and glacial lakes formed in the lowland and in tributary drainage basins. Water in the Champlain lowland flowed south to the Hudson River. Water ponded to the east by ice in the Champlain lowland flowed to the Connecticut River (Figure 2).

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; Figure 2B). Lake Vermont ended when ice-margin retreat allowed the ponded lake water to escape into the Gulf of St. Lawrence. Marine waters entered the isostatically depressed Champlain Lowland (Figure 2C) forming the Champlain Sea. Over the next 1000 to 1500 years, isostatic rebound increased the elevation of the Richelieu River sill at a rate greater than eustatic sea-

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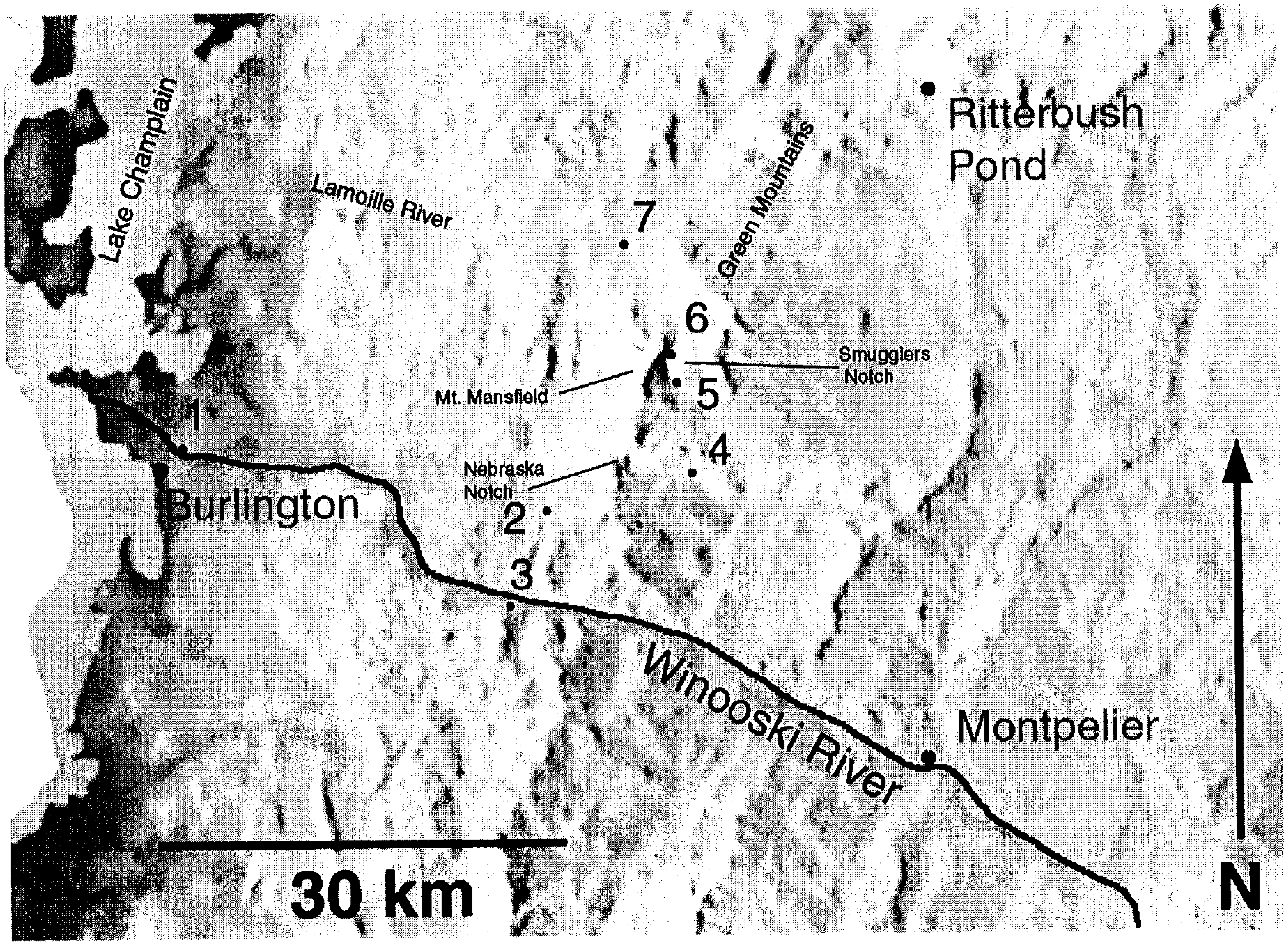


Figure 1. Digital terrain map of north-central Vermont showing the location of field stops discussed in this guide. Sites visited on this trip are identified by number.

evel 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 ınd early Holocene (Figure 3). Valley fills, formed during early, higher base levels, were rapidly incised and left as errace remnants when glacially-impounded lakes drained and base levels not only fell, but were located increasingly arther from tributary valleys (Whalen, 1998). Alluvial fans formed on some of the abandoned terraces, preserving vidence of past sedimentation events. In some locations, these fans archive up to an 8000 ^{14}C yr record of hillslope ıctivity (Bierman et al., 1997).

Slope stability in Vermont is not only a function of natural forcing but also of human activity. Vermont was irst settled by Europeans in the late 1700s. By 1850, much of Vermont had been cleared for agriculture. Initial learance was for cropland; later clearance was for sheep grazing. In the 1870s, land at higher elevations was cleared or timber. Landscape response to this clearance is well preserved in the geologic record. Hillslopes became ınstable and sediment yield appears to have increased significantly (up to 10 times background rates) as documented y aggrading alluvial fans and flood plains (Bierman et al., 1997 and Figure 4).

This trip provides field examples illustrating what we do and do not know about the northern Vermont landscape n terms of deglacial history and slope stability and draws heavily on the past five years of work by students and aculty at the University of Vermont and elsewhere.

ROAD LOG

Mileage

0.0 Begin at the Winooski Mill Parking Lot, Winooski, Vermont. Exit the parking lot from the west side and turn right, heading north, on Routes 2 and 7. Immediately move to the left lane.

0.1 Turn left, heading west, onto Mallets Bay Avenue at the big intersection immediately NW of the Champlain Mill. Road makes a sharp turn to the right in 0.2 mi and continues to north, crossing the railroad tracks.

0.7 Left turn, heading west, on Pine Street.

0.8 Right turn onto Hickok Street at the stop sign at the bottom of the hill.

0.9 Park in cul-de-sac at end of road. Small trails lead NW through the woods to the first stop, Town Line Brook.
UTM Coordinates: 643100, 4928630

STOP 1: TOWN LINE BROOK

Burlington 7.5-minute quadrangle

Geologic Setting

Town Line Brook is a small, deeply incised tributary of the Winooski River (Figure 5). Its valley walls expose ıp to several meters of fluvial gravel cut into tan, well-sorted fine to very fine sand that overlies gray silt with nterbedded fine sand. Underlying these materials are thin silt/clay couplets (rhythmites) near the base of the xposure. The rhythmites were deposited in the quiet waters of Lake Vermont, the surface of which was at an levation of approximately 198 m (650 ft), 143 m (470 ft) above the 55 m (180 ft) terrace that borders Town Line Brook. The overlying gray silt and sand that comprise most of the exposed section were deposited in the Champlain Sea, whose surface lay at approximately 100 m (320 ft), still 45 m (150 ft) above the terrace (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 in this outcrop. The abrupt contact between the ınderlying gray Champlain Sea silt and the overlying fine sand signals the rapid encroachment of the paleo-Winooski River delta, perhaps in response to newly developed distributary channels. Isolated "clay" clumps in the

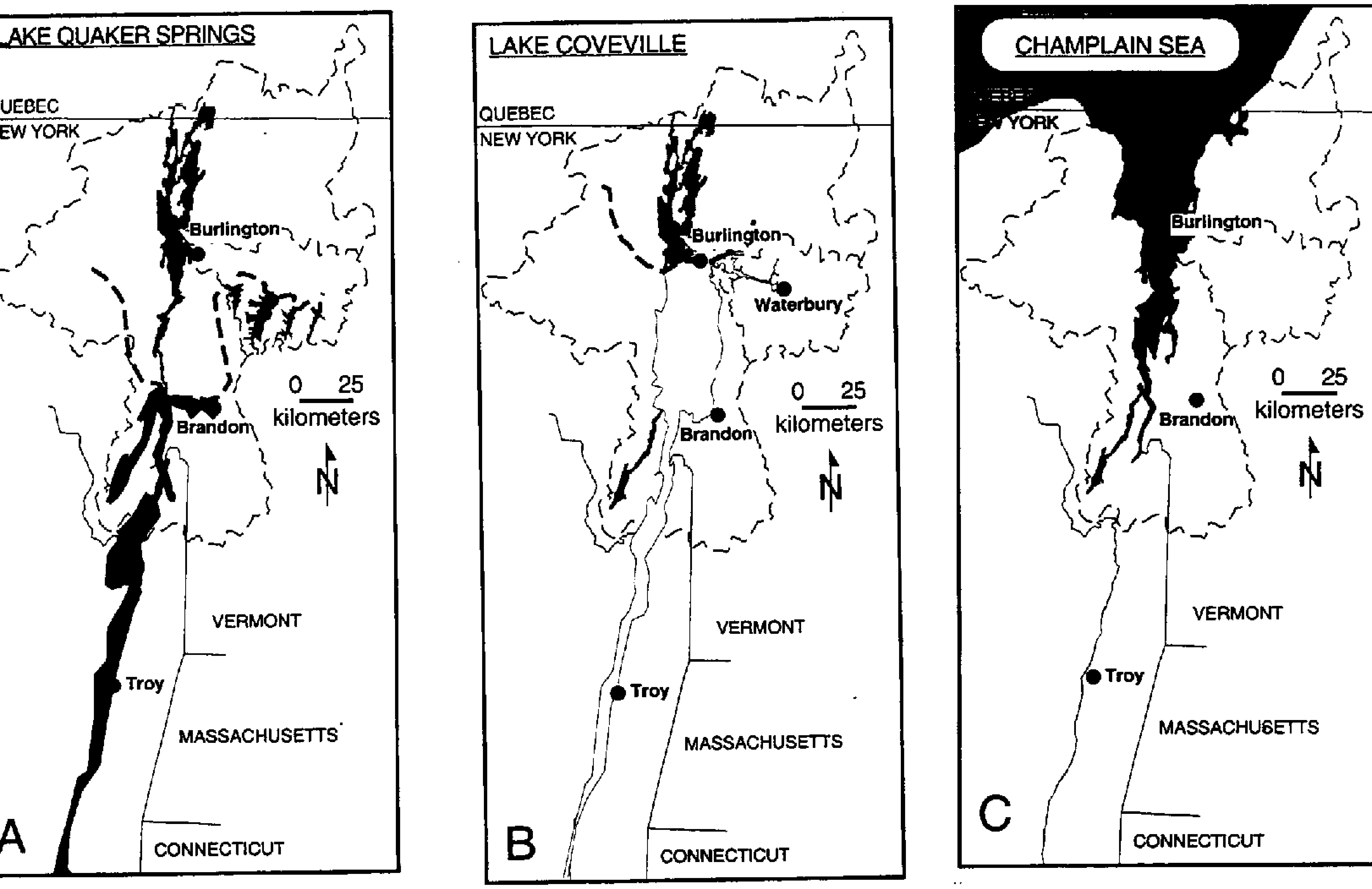


Figure 2. Extent of glacial lakes occupying the Champlain Lowland in comparison with present-day Lake Champlain. A. Lake Quaker Springs, ice margin (heavy dashed line), and isolated ice-marginal lakes in Winooski River Valley. Present-day lake Champlain shown in dark gray. B. Coveville stage of Lake Vermont; ice margin (heavy dashed line) is north of Winooski River Valley and isolated lakes have coalesced into Lake Vermont. C. Extent of Champlain Sea. Diagrams from Whalen (1998; Figures 2.8, 2.9, 2.12).

fine sand indicate that clay deposited in Lake Vermont was being eroded by the river and carried, perhaps in blocks of ice, over the delta before thawing. Coarse sand and well-rounded pebble gravel unconformably overlie the fine sands. Both shallow and deep channels cut into the sand indicate that base level was somewhat lower than the terrace bordering Town Line Brook (55m, 180 ft) at the time the gravel was deposited, indicating that the Champlain Sea had drained and Lake Champlain was no more than 24 m (80 ft) above its current elevation. The north side of the stream (in Colchester) is an old dump sited, as so many are, on the town line!

A perched water table aquifer exists in the fine sand. Groundwater usually seeps from the contact with the underlying gray silt year round. The lower groundwater table intersects the stream and presumably extends up into the silt away from the stream. Storm water drains, located in the development where the cars are parked, may provide a significant additional source of recharge to one or both of the groundwater systems.

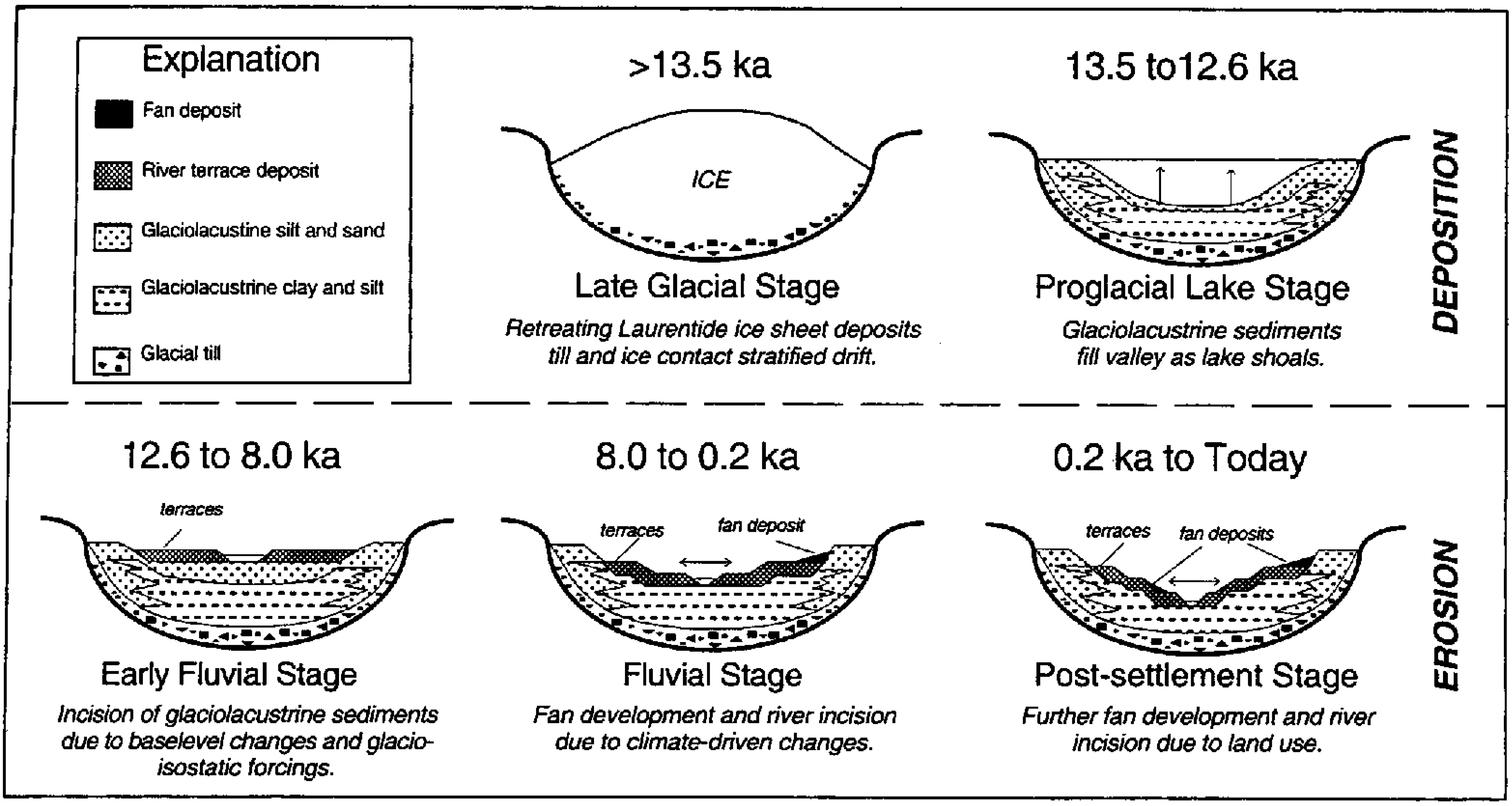


Figure 3. Schematic diagram of valley evolution for the Winooski River drainage basin and its tributaries. Ages are in ^{14}C years BP. From Whalen et al. (1998).

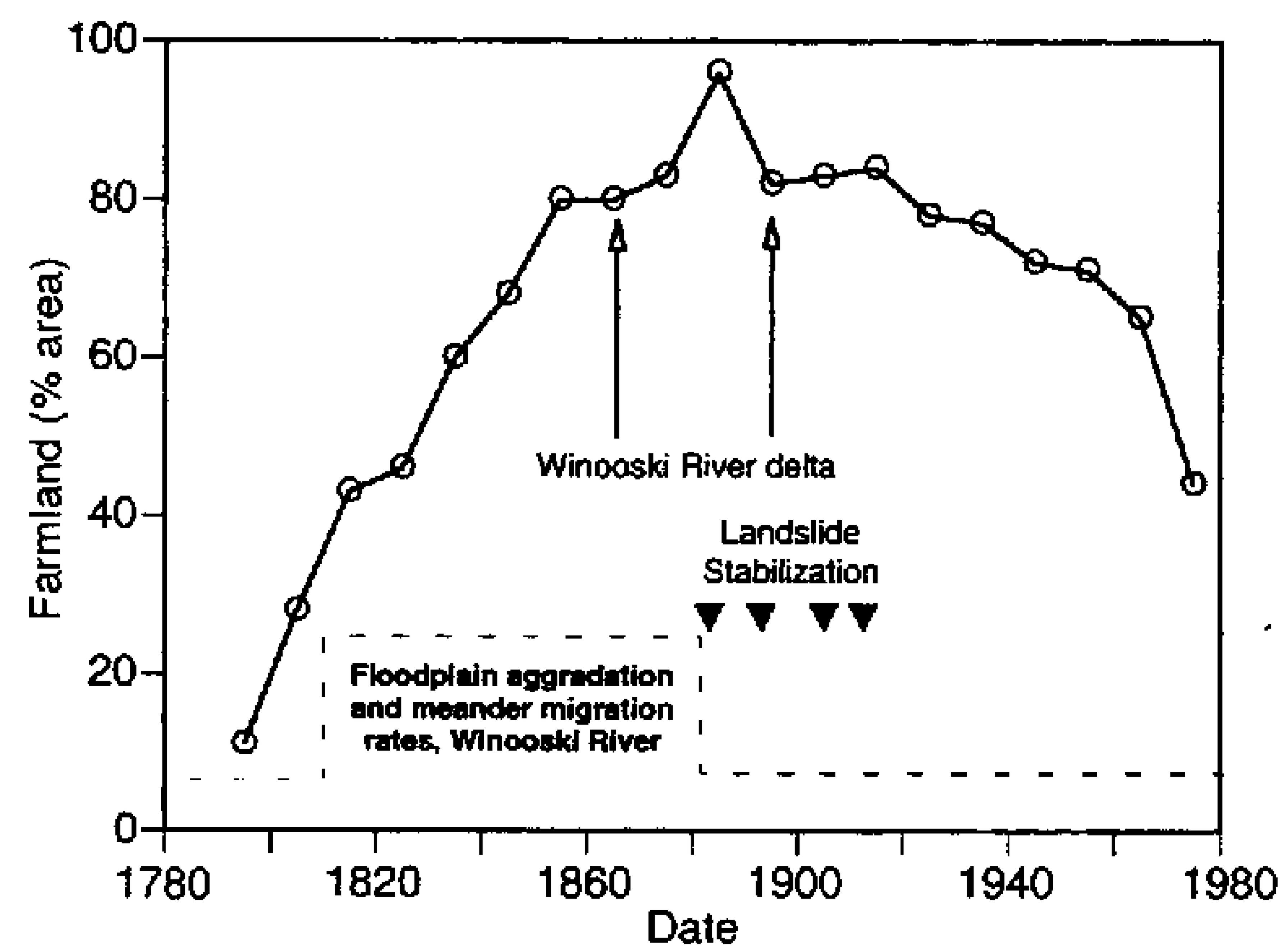


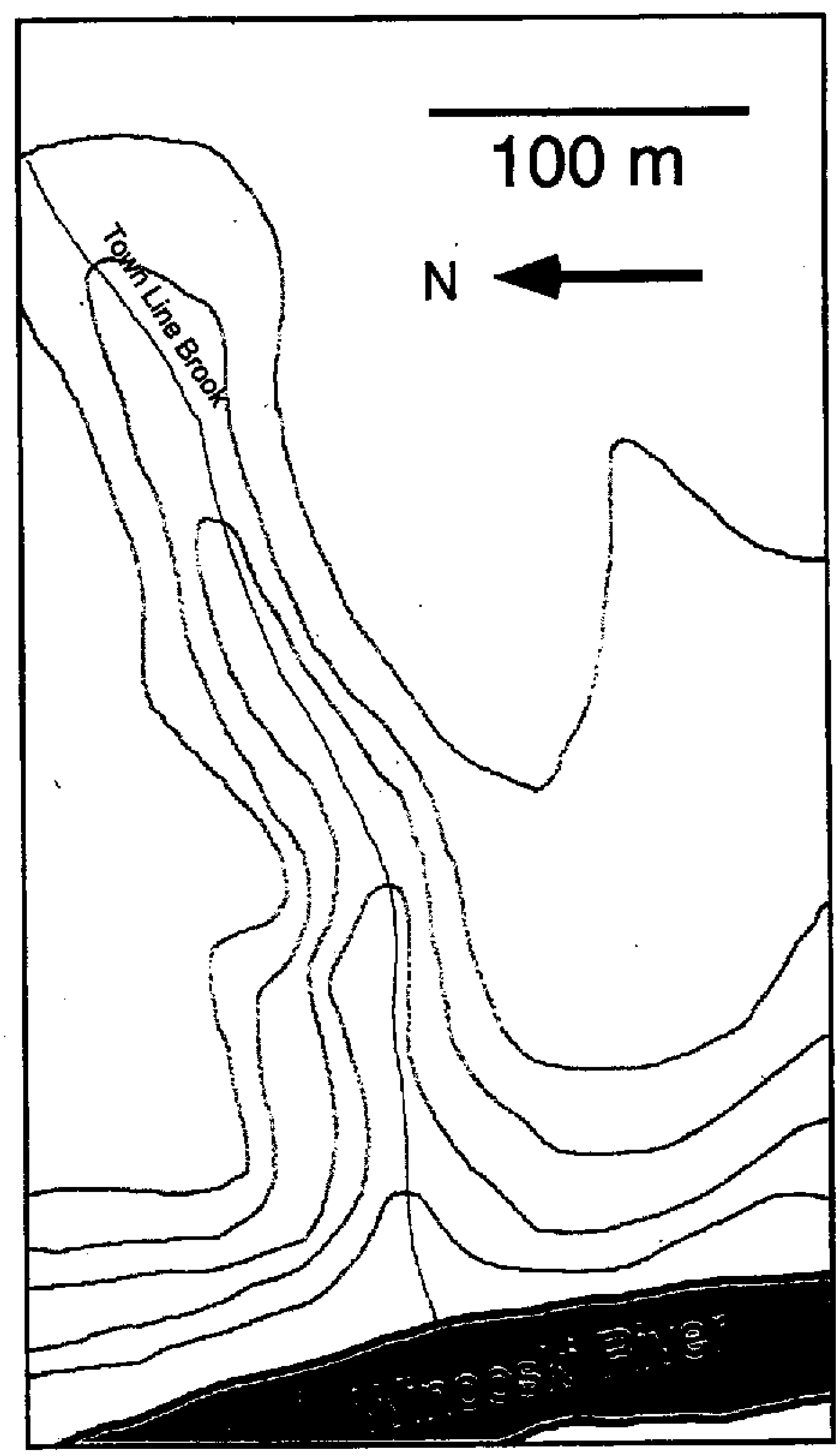
Figure 4. Historic landscape response, Chittenden County, northwestern Vermont. Open circles represent percentage farmland. Major expansion and contraction of Winooski River delta in Lake Champlain as deduced from historic maps are marked by arrows (Severson, 1991). Maximum ages of trees growing in fossil landslide scars on tributary of the Winooski River (Town Line Brook), indicating when slides stabilized, are shown by triangles. Period of increased meander migration and flood plain aggradation in lower Winooski River flood plain shown schematically (Thomas, 1985).

Slope Stability

The valley of Town Line Brook has been and continues to be widened by landsliding and deepened by fluvial incision—controlled to a great extent by the distribution of ground water. Most of the slides that we have observed have occurred by failure of the gray Champlain Sea silts. Typically the slide material consists of, from bottom to top, (1) gray silt that has liquefied and flowed, (2) rigid blocks of cohesive silt, and (3) the overlying non-cohesive fine sand and gravel. The fine-grained Champlain Sea silts contain sandy interbeds along which ground water preferentially flows. We can think of two mechanisms by which the silt fails: (1) The interbeds wash out causing small-scale slumps and toppling failures of the more cohesive, overhanging fine-grained material, which then 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. (2) Alternatively, failure in the silts may initiate in response to high pore-water pressure at the base of the section. 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.

During the last 12 years, slides have occurred one or two times a year along a ~70 m stretch of the south bank of the stream. However, slide scars are prevalent along the watercourse. Ring counting 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 (Figure 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 five years at rates of several cm to >1 m yr^{-1} (Figure 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. These estimates, although crude, imply eposodic landslide activity over the past 10,000 years.

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.



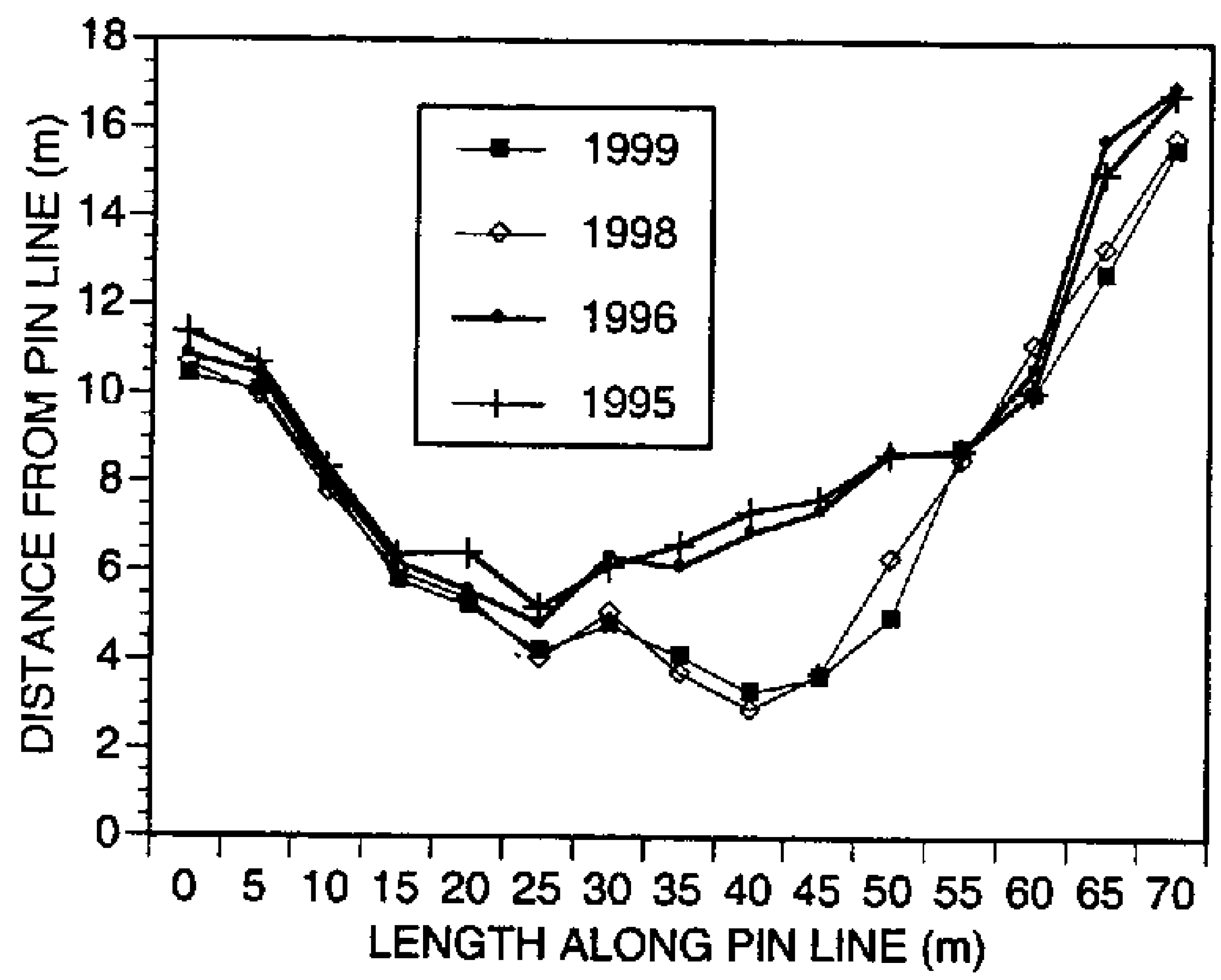


Figure 6. Pin line data for the main Town Line Brook landslide showing retreat of landslide scarp over the past four years. Data gathered by successive UVM Geohydrology classes from 1995 until 1999.

0.9	Retrace route back towards Winooski Mill
1.7	Go straight through big intersection (Routes 7 and 2 with Route 15 and Mallets Bay Avenue) heading east on Route 15.
2.3	Turn right onto entrance ramp of I-89 heading south. Almost immediately the Interstate crosses the Winooski River and offers a brief view of part of the Winooski Gorge to the east, a channel cut in the last 10,000 years after the Champlain Sea drained. The old channel, defined by well logs, cuts NW from Essex, passes under Colchester Village towards Mallets Bay. It is completely buried by Pleistocene sediments. Shortly after crossing the Winooski river, the Interstate turns east and heads for the mountains. Get off at Exit 11, the Richmond exit.
13.6	Exit 11, Richmond.
13.8	Turn right, heading east, on Route 2 at the bottom of the ramp. A nice section of varved Lake Vermont silt and clay was exposed during construction of the Park and Ride opposite this intersection.
15.3	Turn left, heading north up hill at stoplight in center of Richmond Village.
18.4	Bridge over Mill Brook.
19.0	Right turn on Nashville Road heading east. Several kettles have formed in collapsed sand and gravel on either side of the road. Road proceeds east, up the course of the Lee River.
21.5	Bolton Flats, Elev. 238–250 m, 780–820 ft.
22.6	West Bolton cross roads. Go straight through stop sign continuing east on Mill Brook Road. Road is narrow and steep.
23.1	Park where the road makes a turn around loop (shaped like the eye of a needle). **UTM Coordinates: 668140, 4923260**

Continue up the road on foot staying left where the road Y's. Right side of Y is gated and posted. Note old mill foundation between road and stream. Continue until reaching a point where the road abruptly ends at stream crossing (Mill Brook).

STOP 2: MILL BROOK

Richmond 7.5-minute quadrangle

Introduction

Rivers and streams draining the Green Mountains have played extensive roles in Vermont's development providing both water supplies and power, particularly during the 19th century. At this stop we will try to understand the history of one such stream, Mill Brook, that drains westward through West Bolton. Mill Brook, as the name implies, was dammed in several places to supply water to mills (Figure 7). We will be inspecting the site of one such dam.

Geologic Setting

The western flank of the Green Mountains at this latitude is underlain by metasedimentary rocks (dominantly muscovite, chlorite, albite, quartz schists, quartzites, and rare greenstones) belonging to the Underhill formation. The strong foliation and compositional layering in these rocks strike N–S. Streams draining to the west cut across alternately strong and weak rocks. Stream channels are characteristically narrow and contain waterfalls where they cut across the stronger lithologies. The dam site here is one such site. Aside from alluvium, most of the surficial material at this elevation is till although sandy terraces at approximately 408 m (1,340 ft) on either side of the stream suggest that a small ice-bounded lake or stream may have existed here during the retreat of the ice sheet. Landslides initiated by the flash floods of 1990 (described below) yield excellent exposures of gray, clay-rich till. Within this till, blocks of varved silt/clay occur. Given their weakness, it seems likely that the source of these lacustrine sediments was a preglacial lake that may have been trapped between the advancing ice sheet in the Champlain Valley and the mountain side.

Mill Brook Flash Flood, July 1990

Along Mill Brook in West Bolton we will observe the effects of a flash flood that affected many of the drainage basins around Mount Mansfield on July 4, 1990. The weather station on the top of Mount Mansfield recorded 2.10 inches of rain at 4 PM on July 4 and 2.14 inches of rain at 4 PM on July 5th. These readings result from one storm that dumped half its rain before and half its rain after 4 PM in the afternoon when rain collected in the rain gauge is recorded. This is, of course, only an estimate of the rainfall that actually fell in the Mill Brook drainage basin, 9 km SSW of the recording station. This and another flash flood on July 23rd (3.25 inches of rain recorded at 4 PM on the 23rd of July at the top of Mount Mansfield) washed out many roads in the area and significantly altered the course of Mill Brook and many other streams. The drainage basin of Mill Brook above the dam is a bowl covering 6.5 km^2 extending from the main range of the Green Mountains (Bolton Mountain, elev. 1,128 m, 3,700 ft) to the dam (elev. 378 m, 1,240 ft). Using the 4.24 inch (0.108 m) total recorded at the weather station and assuming that this total fell over the entire drainage basin, 700,000 m^3 of water fell on the drainage basin and, given the intensity of the storm and the steep gradients and low permeability soils in the drainage basin (clay-rich till), it is safe to assume that a large percentage of that volume flowed past the Mill site over an unknown period of time, i.e., peak discharges are unknown.

The old (pre-flood) channel of Mill Brook is still clearly visible where it flowed across sand and gravel that filled the former mill pond. It is now possible to determine the sequence of channel jumping events that occurred as Mill Brook repeatedly dammed itself with woody debris (much of this derived from landslides) and cobbles. That sequence will be mapped on the field trip. The final channel-jumping event occurred approximately 300 m upstream from the dam. The newly routed stream presently flows along the south side of the valley. Approaching the dam (now broken and several meters lower than its once functioning height), the stream has eroded its channel through all of the sand and gravel that accumulated behind the mill dam. An old soil horizon is now exposed that contains leaves and other soil detritus, and stumps of trees with clear indications that they have been cut with tools and not chewed by beaver. As a result of the flood, the stream channel has re-exposed the forest floor that existed in the early 19th century when the land was first cleared and the dam first constructed.

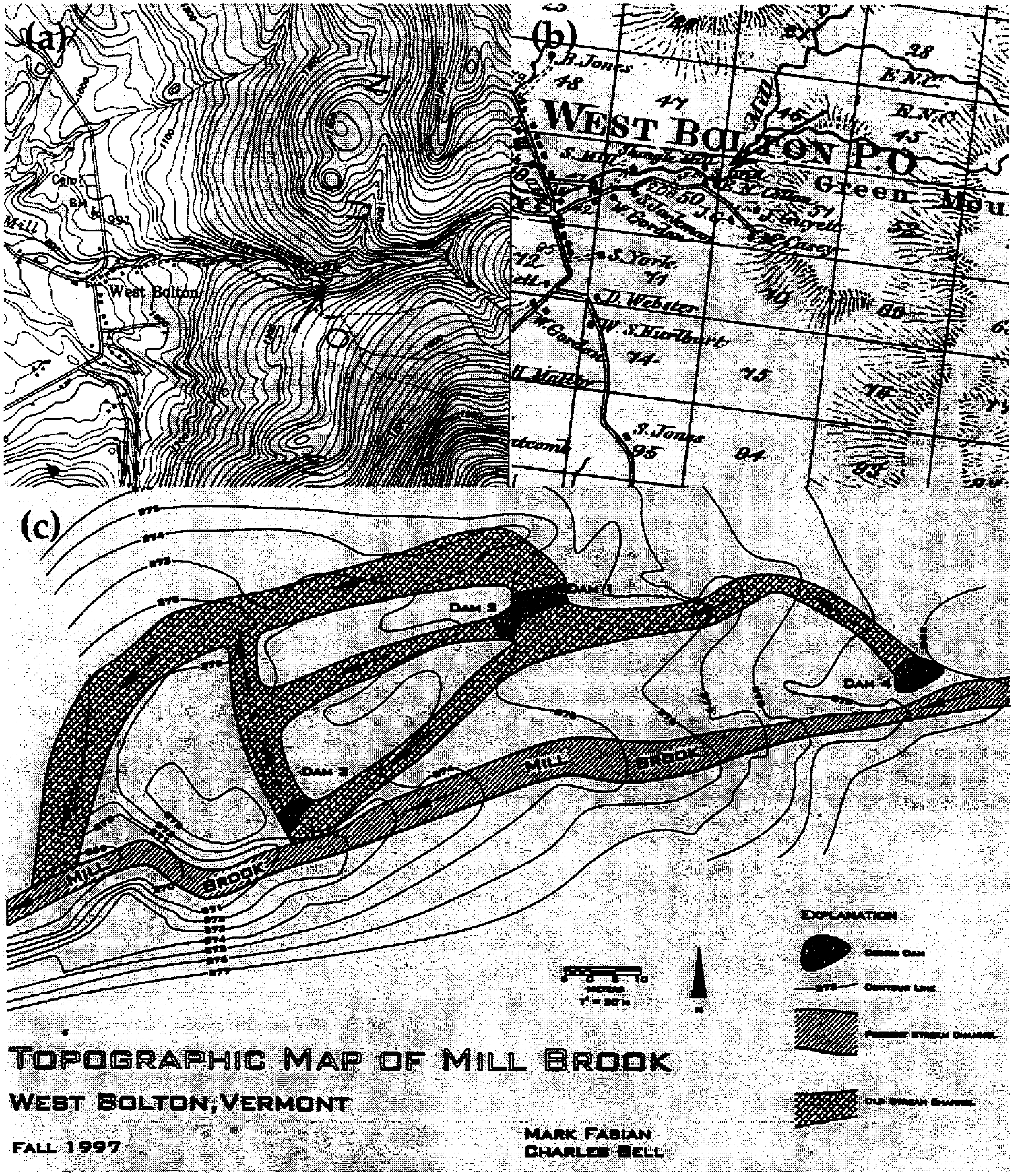


Figure 7. (a) Enlargement of a portion of the Richmond 7.5-minute quadrangle showing West Bolton and part of the Mill Brook drainage Basin. (b) Portion of the Beers Map of West Bolton (1875) showing settlement pattern and mills in existence at the time of publication. Arrows on both maps show location of old dam visited on this trip. (c) Detailed topographic map of the Mill Brook area prepared by students Mark Fabian and Charles Bell. Map shows the different channels that were active during the flash flood, the debris dams that caused those channels to jump, and the present course of Mill Brook.

23.1 Return to cars and head back down road to West Bolton.

23.6 Turn left, heading south, on Stage Road.

23.9 Turn left on Bolton Notch Road (West Bolton Golf Course is on the right).

24.7 Drainage divide in Bolton Notch, elevation 366 m, 1200 ft.

27.5 Gravel pit on west side of road is cut into a large delta with south-dipping foreset beds. Present terrace above gravel pit is at an elevation of 232 m, 760 ft, approximately 31 m, 100 ft higher than the elevation of Glacial Lake Vermont at this latitude.

28.2 Turn right (west) on Route 2.

29.1 Jonesville: Turn left onto the single-lane steel bridge over the Winooski River. The Huntington River joins the Winooski River just downstream from the bridge (west). A large gravel bar deposited by the Huntington River has accumulated where the two streams join.

29.3 Turn left, heading east, on the Duxbury Road.

30.0 Park along side of road next to large, glacially polished and plucked outcrop on the south side of the road.
UTM Coordinates: 665220, 4915880

STOP 3: JONESVILLE ROCK
Richmond 7.5-minute quadrangle

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 function 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 (Figure 8).

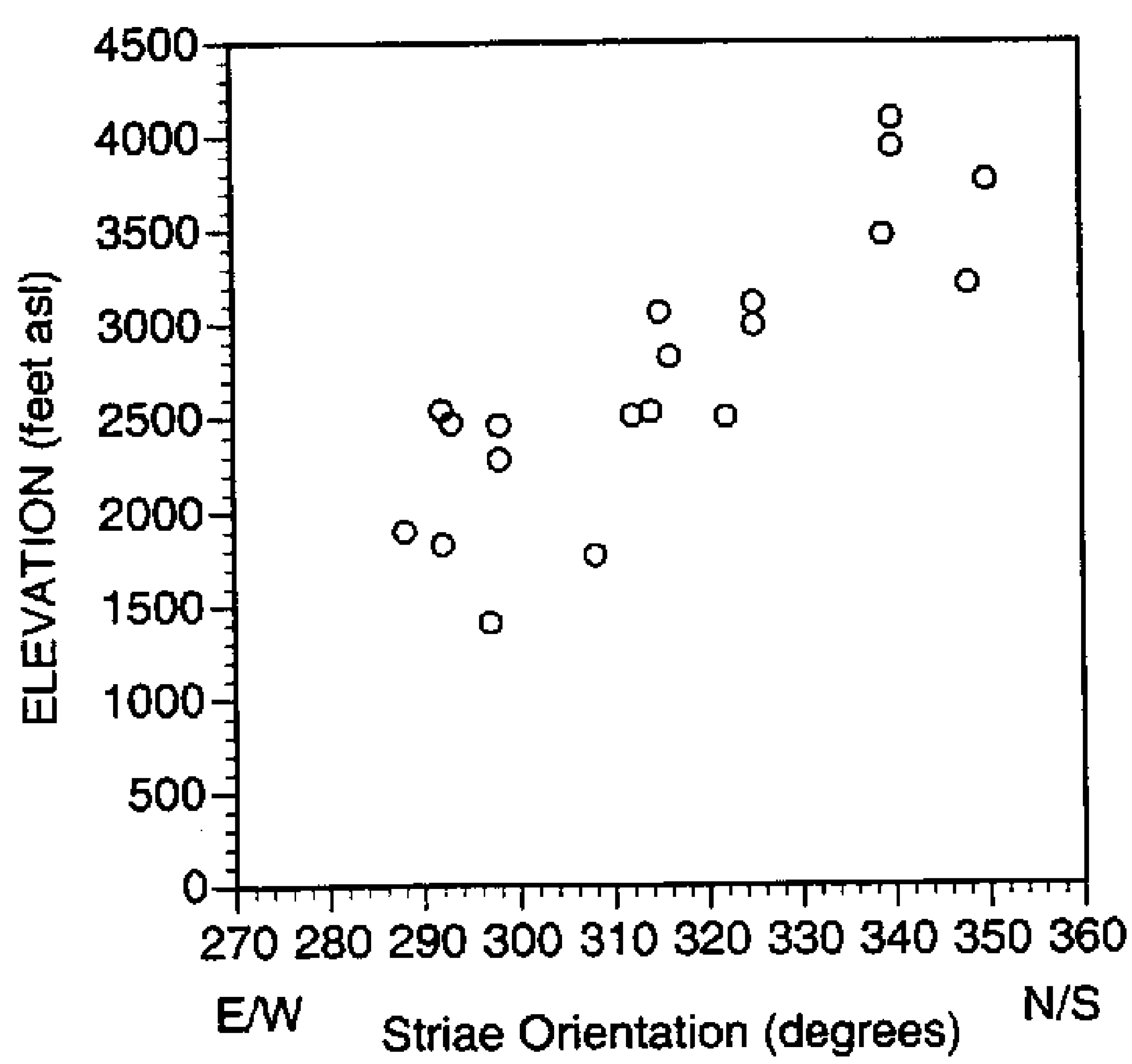


Figure 8. Orientation of striae as a function of elevation, transect from near Jonesville Rock to the summit of Camels Hump; figure from Malchyk and Kelly (1996).

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 (Figure 9), 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.

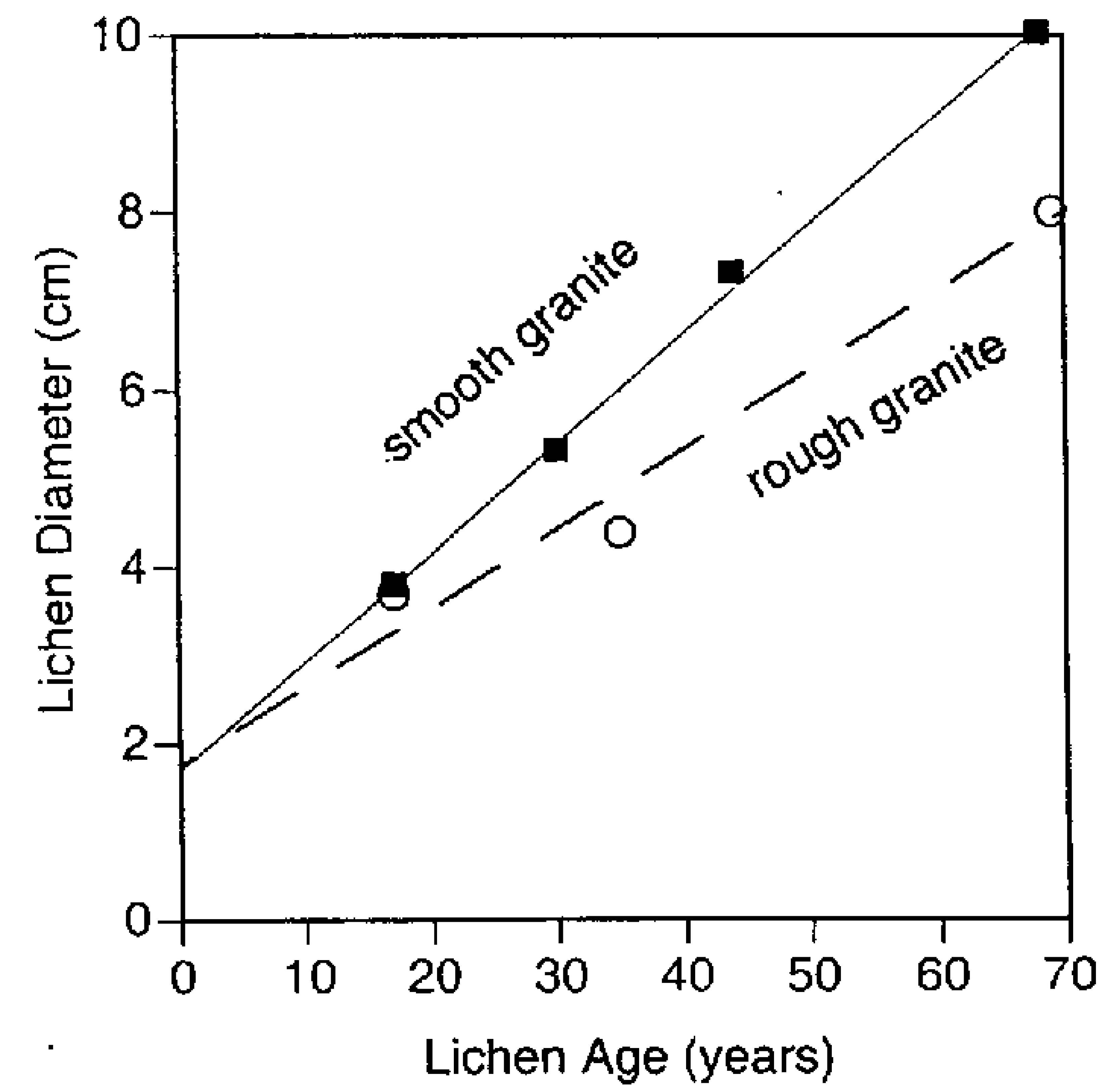


Figure 9. Calibration of lichen (probably *Xanthoparmelia plittii*) maximum diameter for the Richmond, Vermont area (Sundae, 1997).

OPTIONAL SIDE TRIP TO HUNTINGTON GORGE
(Road log not included)

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 N82° E, N75° W, and N51°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, ^{14}C-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 (Figure 10). The T6 terrace was graded to the Fort Ann stage of Lake Vermont, which ended 11,700 ^{14}C y BP with the initiation of the Champlain Sea. Thus, 11,700 ^{14}C y BP, is the upper limit for the age of the gorge. The gorge may have been exposed as late as 8500 ^{14}C 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 (Figure 10). These dates lead to an important conclusion; the gorge was not formed by fluvial erosion related to the latest deglaciation nor was it formed by the catastrophic draining of the last glacial lake to occupy the Huntington Valley. Either the gorge was formed by erosion through the Holocene or the Huntington River excavated and reoccupied a gorge that was cut previous to the last glaciation.

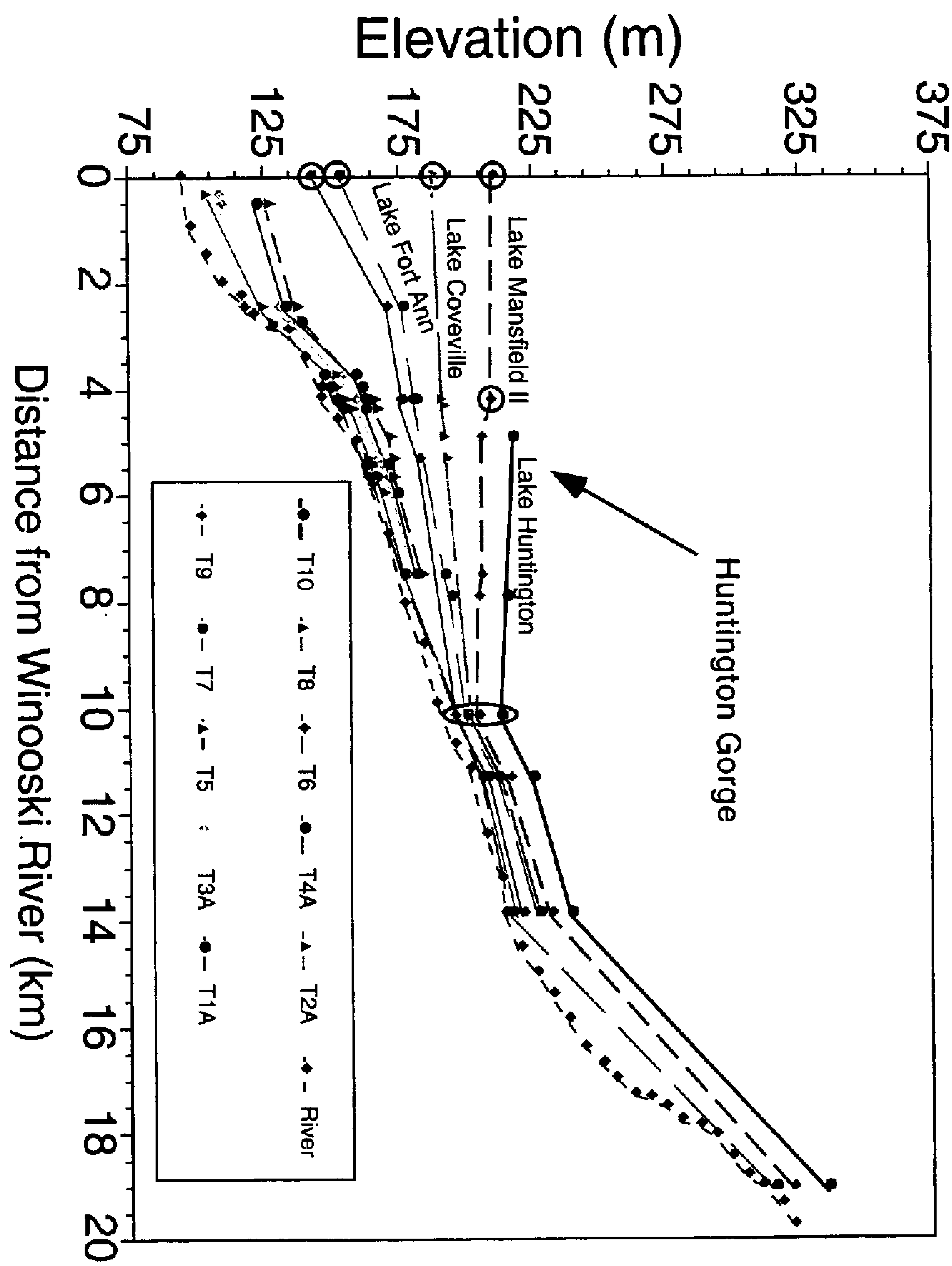


Figure 10. Longitudinal profile of terraces bordering the Huntington River Valley. Adapted from Whalen (1998, Figure 4.2). Figure is vertically exaggerated.

30.0 Retrace route back over bridge to Route 2.

30.9 Turn right, heading east, on Route 2. Follow this, through Bolton, to Waterbury.

39.8 Intersection with Route 100. Turn left, heading north, crossing over the Interstate and continuing to Waterbury Center.

42.2 Well developed terrace (bedrock core!) in pasture on west side of road is at an elevation of 204 m, 670 ft—coincident with the elevation of Glacial Lake Vermont.

43.5 Cider Mill, Waterbury Center: Lunch Stop. Continue north on Route 100.

47.5 Turn left, heading west, onto the Moscow Road. Sign also indicates that this is an alternate route to the Mount Mansfield Ski Area. Continue through the village of Moscow.

49.0 Barrows Road intersection. Continue straight (west).

49.6 Bridge over Miller Brook. Follow road around to right, now called the Nebraska Valley Road.

50.0 Look for white house (former barn) on left side of road and continue another 100 m and park in meadow on left side of road adjacent to an active alluvial fan.
UTM Coordinates: 679300, 4923650

STOP 4: MILLER BROOK ALLUVIAL FAN AND GULLY

Stowe 7.5-minute quadrangle

Introduction

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 and irregularly 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.

Geologic Setting

Almost 6 m of lacustrine sediments are exposed on the near-vertical walls of the gully above the fan, extending from the top down to the contact with the underlying till. These sediments were deposited in Glacial Lake Winooski, a lake dammed by ice in the Winooski River valley whose outlet lay approximately 4 km south of Williamstown (see further descriptions of this lake in both Wright, 1999 and Larsen, 1999, this volume). At this point, the lake surface was at approximately 337 m, 1,100 ft (134 m, 440 ft above the gully). This section, deposited directly on till, records the early sedimentation history in this part of the lake when the ice front was probably quite close by. Most of the sediments accumulating in this part of the lake were derived from the mouth of an esker tunnel that lay at the base of the ice sheet and extended at least as far up valley as Lake Mansfield (Wright et al., 1997).

The bottom of the section consists of a thin (0.1–0.2 m) layer of coarse sand and pebble gravel deposited on top of the steeply dipping till surface. Above this (from 3.1 down to ~5.5 m; datum is top surface of the gully) is a highly disturbed (slumped) section of lacustrine fine to very fine sand, silt, and clay. Bedding, where visible, is folded and pockets of sand are completely surrounded by clay. The slumped sediments below are overlain by a 1 m thick section (from 2.1 down to 3.1 m) of horizontally layered, medium to fine sand/silt and clay couplets (couplets are 10–15 cm thick). The sand in many of these layers has been severely disrupted by soft sediment deformation occasioned either by loading or by seismic activity. Above this lies 1.3 m (from 0.8 m to 2.1 m down) of finely laminated lacustrine silt and clay with minor very fine sand. The top 0.8 m of the section consists of pebbles suspended in a structureless matrix of very fine sand and silt. It is unclear whether this material is (1) a debris flow deposit, (2) a particularly dense accumulation of dropstones, or (3) fill, perhaps added at some point in the past when the gully first started to form.

Fan Dynamics

A long (~50 m) narrow (~6 m) gully supplies sediment to the fan (Figure 11). 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. The pipe(s) are eroded from the fine lacustrine sand and follow the steeply dipping contact between the relatively impermeable till below and the very permeable sand and gravel horizon. 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 ($< 1.5 \times 10^{-3}$ 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.

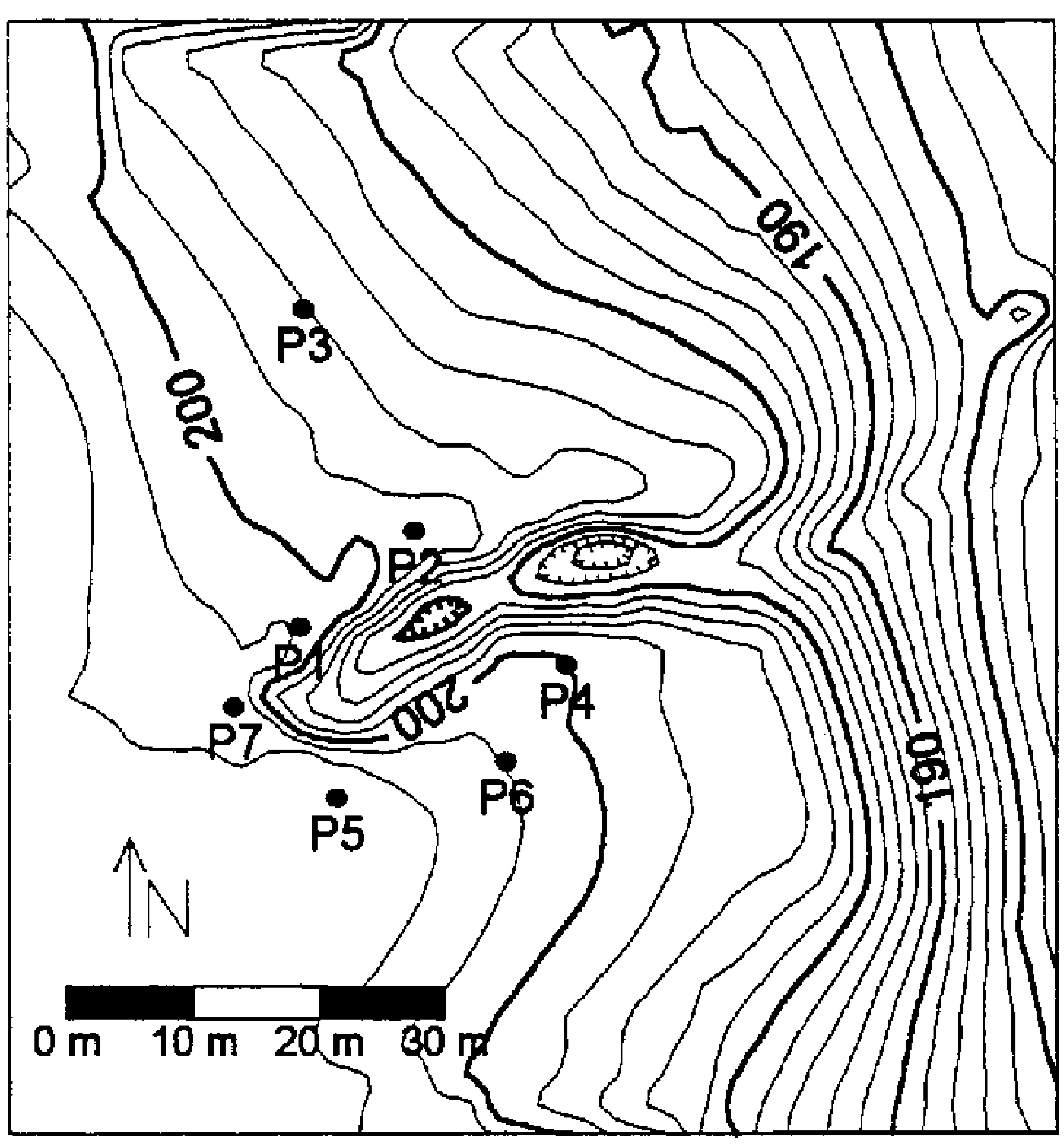


Figure 11. Map of gully and terrace immediately above active Stowe alluvial fan. Map produced using Pentax Total station and Trimble GPS 4400 RTK system by University of Vermont, 1998 Geohydrology class.

We have only recently identified the process that most likely led to the formation of this gully. Neighbors believed 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). Further erosion of the pipe in the wet summer of 1998 prompted the roof of the pipe to collapse. This collapse revealed that the pipe was formed in a lens of sand and gravel within the fine sand and silt. The high hydraulic conductivity of the sand and gravel compared to the lower hydraulic conductivity of the fine-grained lacustrine sediment focused water flow and promoted erosion. If this hypothesis is correct, the natural piping system eroded the gully from the bottom up rather than the more-familiar top-down erosion of surface drainage networks.

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.

50.0 Retrace route back to Barrows Road Intersection

51.0 Turn left, heading north, on Barrows Road.

52.7 Turn right on Luce Hill Road. Road to left goes to the Trapp Family Lodge.

53.3 Turn left, heading northwest, at intersection with Route 108.

56.8 Road pitches upward steeply as it climbs the front of a sand and gravel deposit that is probably a delta. Terrace opposite Mount Mansfield Cross-country ski center is at an elevation of 357 m, 1,170 ft. This falls on the projected water surface plane of Glacial Lake Winooski (see discussions elsewhere in this volume by Wright, 1999 and Larsen, 1999).

58.7 Ski slope entrance on left.

60.1 Pull over along wide margin opposite toe of debris flow, now partially covered with young vegetation.
UTM Coordinates: 675340, 4934560

STOP 5: SMUGGLERS NOTCH DEBRIS FLOW

Mount Mansfield 7.5-minute quadrangle

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, approximately 250,000 m^3 of material, originated in the gully extending up the east side of the valley below Spruce Peak and incorporated colluvium as well as trees and soil (Lee et al., 1994). An intense rainfall apparently initiated 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 accumulate sufficient debris so as to again present a hazard.

60.1 Continue northwest up to top of Smugglers Notch.

60.4 Big Spring on right.

61.1 Parking Lot at top of Smugglers Notch.
UTM Coordinates: 675100, 4935800

STOP 6: SMUGGLERS NOTCH AND MOUNT MANSFIELD

Mount Mansfield 7.5-minute quadrangle

Introduction

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).

Geologic Setting

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

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 is described by Baskerville et al. (1988). The landslide began at about 7 a.m. when a large block of rock ($\sim 10.4 \times 10^6$ 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.

Mount Mansfield

Mt. Mansfield, at 1340 m (4393') is the highest point in Vermont. It is one of only five peaks in Vermont that exceed 1220 m (4000') in elevation, the others being Killington Peak at 1293 m (4,241'), Camels Hump at 1248 m (4,093'), 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 (Figure 8).

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 12). Winter inversions reduce effective 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.

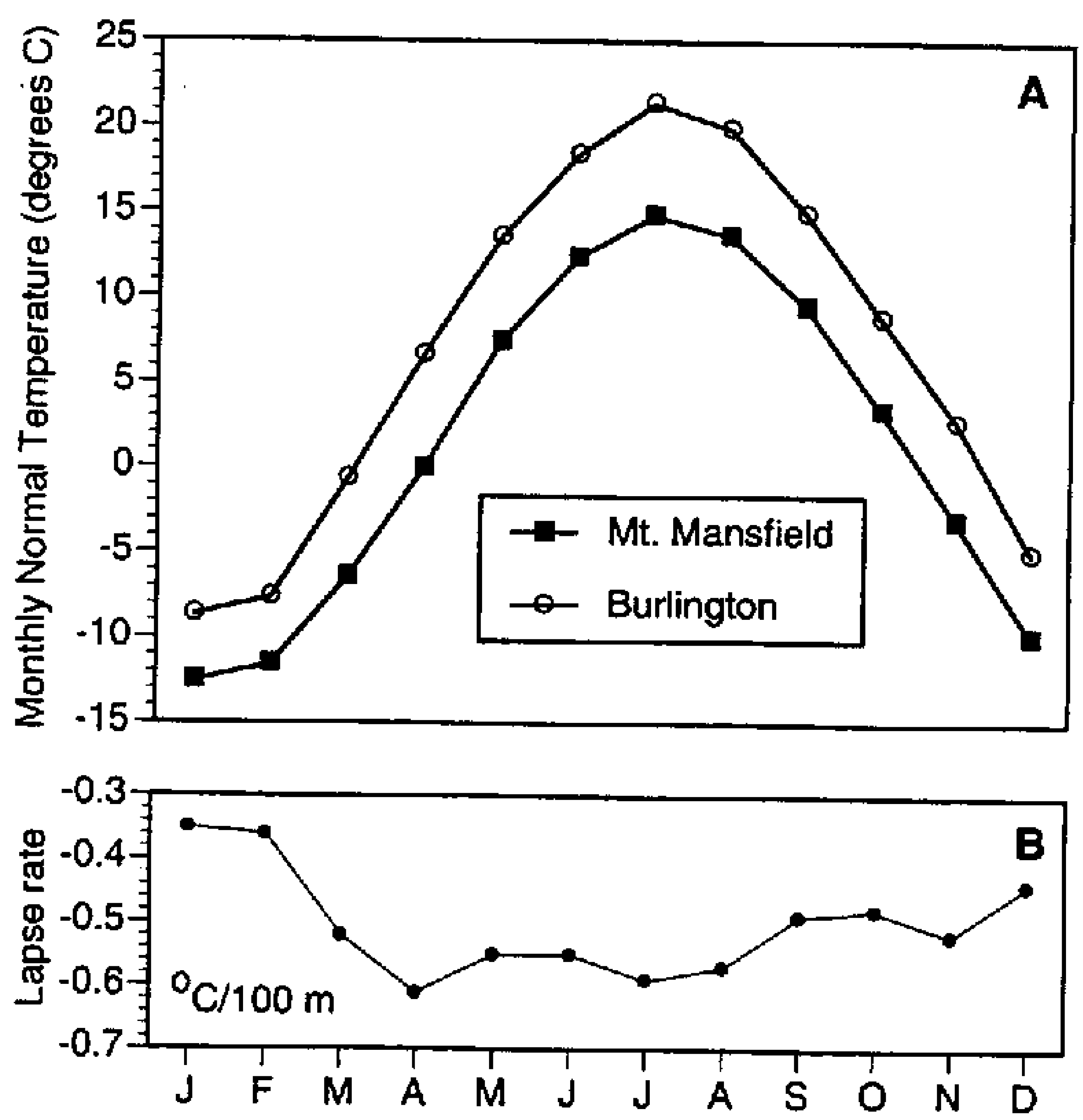


Figure 12. Monthly average temperatures and lapse rates for Mt. Mansfield weather station (1204 m asl) and Burlington weather station (101 m asl). Data from National Weather Service. A. Monthly average temperatures. B. Effective monthly average lapse rates.

61.1 Continue north, down Smugglers Notch.

62.7 Upper lift area for Smugglers Notch Ski Center.

64.0 Smugglers Notch Village

66.9 Gravel pit in 244 m, 800 ft terrace consists of coarse sand and pebble gravel comprising topset and foreset deltaic facies.

68.9 Village of Jeffersonville. Turn right (north), in front of hardware store, onto Main Street.

69.0 Turn right, heading east, on School Street.

69.2 Park in the lot behind the Cambridge Elementary School.
UTM Coordinates: 672380, 4945560

STOP 7: JEFFERSONVILLE CLAY BANK: 1999 LANDSLIDES

Jeffersonville 7.5-minute quadrangle

Introduction

The Jeffersonville slides of 1999 are a series of three mass movements that have generated a dramatic scarp and attracted significant local media attention. The slides have undermined the foundation of one house and currently threaten several others nearby (Figure 13). The three slides disrupted a 50-m high stack of glaciolacustrine material, carrying more than 27,000 m^3 of silt and sand over a convex, cut bank of the Brewster River and onto the point bar and low terraces on the other side.

The debris was unusually mobile, moving more than 150 m laterally on what geologic evidence suggests was a rapidly deforming, fluidized bed. The landslide probably owes its long run out distance to high basal pore pressure, the best evidence for which is a series of fluid escape structures, mud volcanoes. Excess pore pressures, in the remobilized fine-grain material, allowed some of the slide material to flow as a non-Newtonian fluid forming distinct debris-flow like snouts at the margin of the run out zone.

Geologic Setting

The village of Jeffersonville is built on a large alluvial fan deposited by the Brewster River as flow became unconfined and as its gradient abruptly lessens where it meets the Lamoille River floodplain. The Brewster River channel is currently incised into the east side of the fan and has remained in that position for at least the last 122 years, (Beers Atlas map of Jeffersonville, 1877). A large rooted tree stump was exposed at river level by minor flooding approximately 5 years ago. While we did not date the stump, it nevertheless indicates that the fan has aggraded during the Holocene, similar to other alluvial fans in northern Vermont (Bierman, et al., 1997).

The Jeffersonville slides originated from a steep bank (the "Jeff Clay Bank") of unconsolidated glacial and immediately post-glacial sediments. The bank is the western margin of a North-South oriented ridge defined by the Brewster River to the West and an unnamed stream to the East (Figure 13). The top of the ridge is flat, most likely a fluvial terrace of the ancestral Brewster River as it began eroding the thick section of glaciolacustrine material exposed in the landslide scarp.

A detailed stratigraphic section was measured by one of us (SFW) during the summer/fall of 1991 at the site of the current slide (Figure 14a). The measured section begins approximately 2 m above water level and extends to within 15 m of the top of the bank. The entire measured section (29 m) consists of 143 couplets (rhythmites) consisting of varved silt/clay that grade into fine sand/clay couplets higher in the section. The silt layers in the lower part of the section range in thickness from 6.7 to 51.6 cm and the clay interbeds range from 0.4 to 2.9 cm. The clay layers are usually graded, becoming progressively finer grained from bottom to top. These are relatively deep water lacustrine sediments deposited in one of the lakes that occupied the Lamoille River valley once the ice sheet had retreated NW of this point (Glacial Lakes Mansfield and Vermont; Connally, 1972). An unknown thickness of surficial materials (most likely lacustrine clay and till) lies below the exposed section. A similar section was measured by Antevs (1922, his section number 170) approximately 50 m to the north, although he only measured the silt/clay couplets comprising the lower half of the section.

The clay layers are systematically jointed into orthogonal sets (Figure 14b). Typically the clay layers break into rectangular pieces along these joints. At present it is unclear whether the joints have formed in response to a regional stress system in the last 13,000 years or whether the stress system is local, perhaps resulting from gravitational stresses along the steep bank. The joint surfaces are frequently stained red or orange with iron oxide minerals suggesting that groundwater flow, at least near the surface, is controlled by this secondary porosity in the clay layers.

Two massive slumps (underwater landslides) in the bottom part of the section transported considerable thicknesses (0.81 and ~5.5 m respectively) of lacustrine clay and sand to this part of the lake bottom (Figure 14).

The upper slump contains an impressive array of deformation structures including abundant folds and imbricate thrust faults. Both slumps are conformably overlain by quiet water silt/clay couplets indicating that normal sedimentation resumed after the slumps. It is presently unclear whether these slumps were initiated by seismic activity, storms, oversteepened slopes, or some combination of the above.

Higher in the section fine to very fine sand becomes increasingly abundant and completely replaces the silt component of the couplets (Figure 14a and 15). These are most likely bottomset beds of a delta built into the glacial lake by the Brewster River, perhaps 1.5 – 2.5 km south of the clay bank. An unconformity separates the fine lacustrine sand and clay couplets from fluvial gravels that comprise the upper ~2–5 m of the section (not measured in Figure 14a). These were probably deposited by the Brewster River after the glacial lake drained and represent an early part of the history of erosion of the lacustrine materials that continues to the present day.

The entire section of lacustrine material visible at the Jeffersonville clay bank was probably deposited in a relatively short period of time. In addition to the 143 couplets measured in the section, the bottom 2 m of covered section probably contains another 15 varves (extrapolating the average varve thickness down section). In addition, the top, unmeasured ~15 m, part of the section may contain another 10 –15 sand/clay couplets below the unconformity with the overlying gravels. If one interprets each couplet to represent a year's sedimentation cycle in the lake, then the measured section was deposited in 143 years and the entire exposed section in ~170 years. This rapid deposition was no doubt influenced by the close proximity to the Brewster River delta noted above.

Slide history

The Jeffersonville slide appears to be one of the largest mass movements in the collective memory of living Vermonters. On April 11, 1999, the first of three landslides occurred at this site. A week later, (April 18), the largest of the three slides released. On July 4, 1999, a third slide occurred at the same location. The slides garnered significant media attention because they progressively endangered a house at the scarp margin and because two houses in the run-out zone were affected by the second failure (mud splash only). Residents were concerned about the possibility of future slides because the debris could dam the Brewster River, a potentially hazardous situation for village residents, all of whom live on the alluvial fan created by the Brewster River.

The Jeffersonville clay bank has been the locus of landslide activity for quite some time. Antevs (1922) measured a section of varves 50 m downstream of the present slide suggesting that exposure was good then, probably from a recent slide. Our own observations over the last 12 years indicate that much of the clay bank has been exposed by small-scale slumps during all of that time. Reconnaissance mapping up- and down-stream of the slide, as well as interviews with local residents, suggests that similar failures have happened here before. As one faces the scarp and looks across the river, there is a well-vegetated slide scar just upstream of the current scarp, probably the result of a similar but smaller slide that occurred in the 1950s. This slide also crossed the Brewster River. Local accounts suggest that the fine grain material from the 1950s slide was used to create the clay tennis courts that were largely buried by the 1999 slides.

Immediately after the second slide, there was public insistence that some type of remediation be conducted to prevent future slides and lessen the possibility of a future slide damming the river. The state of Vermont, employing a local contractor, spent $40,000 "cleaning the channel" below the slide for a distance of 175 m during which time small caliber (< 1 m to 1.3 m) rip-rap was installed 2 m high on the slide-proximal bank of the river. The third slide ran over the rip rap, burying some blocks and carrying other blocks over the river and up the debris apron on the distal side. After the third slide, bulldozers were again used to clean the channel but no further rip rap was applied. The State of Vermont is currently studying remediation alternatives and the Vermont Agency of Natural Resources has concluded that the preferred option for addressing issues created by the slide is removal of all slide material from the 100-year flood plain (virtually the entire debris apron) and extension of the rip-rap another 70 m down stream to prevent oversteepening of the bank by lateral migration of the Brewster River.

Slide Morphology

The slides originated from a 50-m high bluff, the steepness of which is maintained by the Brewster River, which currently erodes the toe of the slope. Observations, made prior to the 1999 slides, indicate that the slope was bare and free of vegetation from the stream up to the top of the slope (Figure 13b). The old nonvegetated slide scarp was relatively narrow (~20 m across) on the upper part of the slope (>15 m above river level), but widened at the base (from river level up to ~15 m) to least 100 m. The exposed area was much smaller than the present scarp, but evidence of small-scale, recurring landslides was abundant. The break in slope, marking the top of the lower slide, occurs at the top of the massive 5.5 m thick slump that is interbedded with the varved silt and clay (Figure 14). Areas both north and south of the upper part of the preexisting scarp were tree covered, but were undercut by the active slides on the lower slope.

The current slide scar is arcuate in plan view and is about 150 m long. A steep upper section extends down from the upper terrace to a distinct bench in the lower section. The bench occurs at the level of the massive slump (Figure 14) and has a steep margin on the side of the Brewster River. Approximately half way down the slope is a cluster of back tilted trees transported down slope from the terrace top. To the north of the active slide, the silt is exposed at and near the river level in a steep bank. To the south of the current slide area, there is an older, well-vegetated arcuate slide scar separated from the current failure by a short ridge. Farther south (upstream), the lacustrine silt appears to be underlain by till (L. Becker, personal communication), an observation supported by the frequency and size of boulders in the Brewster River Channel.

The run out zone was littered with downed trees. Mapping done just after the second slide indicated that most of the trees were concentrated at the perimeter of the slide (Figure 16). The orientation of the trees in the run out zone with respect to the nearest debris margin was bimodal, parallel and perpendicular to the direction of run out. Perpendicular orientations are consistent with trees pushed in front of the moving debris. Trees oriented parallel to the transport direction presumably moved on the debris.

On the basis of a detailed GPS survey (>3000 points) conducted within days of the second slide, we estimate that the volume of the run out material on the west side of the Brewster River was about 23,000 m^3 of silt, sand and gravel (Figure 17A). This volume estimate is somewhat sensitive to assumptions regarding initial topography and could be 20 to 30% larger if the flood plain elevation were significantly lower than the elevation of the flat areas just distal to the run out margins. The volume of material crossing the Brewster River during the smaller third slide (4200 m^3 not including additional material removed by bulldozers in the stream channel) was estimated using a second GPS survey in early July (Figure 17B).

Hydrologic Effects on Slope Stability

The drainage basin above the slide is small and narrow. Beyond the ridge visible from the slide debris, the slope quickly falls off on the other side toward an unnamed brook (Figure 13). Nevertheless, during wetter times of the year, seeps emanate from the top of the silt just below the region where the material becomes sandier. Similar stratigraphic control on groundwater flow is well demonstrated at other sites such as Town Line Brook.

Several lines of evidence suggest that the varved silt and clay at the base of the section was saturated when it failed. The silt and clay adjacent to the river channel is usually saturated and regularly slumps or flows into the river channel. The material flowed immediately after failing in the slide. Dewatering of the slide material took days to weeks. Saturation of the fine-grain material prior to failure would have lowered resisting forces (decreased normal force on the failure plane) and increased driving forces (increased slide mass).

Curiously, the fall of 1998 was much drier than normal; the winter of 1998-1999 was somewhat drier than normal and the spring and summer of 1999 were much drier than normal (Figure 18). In fact, the summer of 1999 was the fifth driest on record. However, 1998 was the wettest year on record in Burlington, Vermont with annual precipitation 50.42", 152% of the average amount. The summer of 1998 was particularly wet. Burlington received 24.77" of precipitation in June, July and August, surpassing the previous record by more than 2 inches.

Although it is not possible to link the slide directly to moist antecedent conditions, it is possible that the wet summer and fall of 1998 may have helped induce the failure. Higher than average summer flow in the Brewster River undoubtedly occurred in response to unusually heavy summer rains in 1998; these flows likely undercut the bank directly below the slide, the outside bend of a meander. It is also possible that the low permeability of the silt retarded drainage of water infiltrated during the previous summer, maintaining high pore pressures in the silt as a result of the heavy rains of 1998.

Larry Becker, Vermont State Geologist, has proposed another means by which to increase pore pressure in the underlying silt. The bed of the unnamed stream to the east of the active slide is 10 to 12 meters above the Brewster River (Figure 15). Movement of groundwater exfiltrated from the stream toward the Brewster River could substantially raise pore pressure in the silt, reducing resisting forces.

Slide Dynamics

Observations made just after the second failure, some of which can still be made today, five months after the event, clarify slide dynamics. In particular, field observations allow us to constrain the type of failure and the mode of debris transport.

It appears from the stratigraphy preserved in the debris apron that the failure was, at least in part, translational. Closest to the river and to the slide scarp, most material exposed at the surface of the slide apron is sand. Moving away from the river, the distal debris apron is increasingly dominated by silt and fine sand. This spatial pattern suggests that the failure initiated in lower part of the scarp (the glacial-lacustrine silt) which rotated toward and then over or through the river. The sand, which is stratigraphically higher on the scarp, then followed in translation. The overall stratigraphy preserved in the run out zone, silt farther from and sand closer to the slide scarp, is consistent with such a translational mechanism as opposed to the toppling failures frequently observed at Town Line Brook (Stop 1).

The style of failure and the mobility of the material have important public policy and management implications. The mobility of the failed material allowed significant run-out putting residences over 150 m away from the slide in harm's way. However, the same mobility allowed the slide to translate away from its source leaving mostly fine sand (easily erodable material) in the channel of the Brewster River. Observations of mudlines made just after the second slide indicated that river stage up stream of the slide remained below bankfull, mandating that the river very quickly cut down through whatever slide material (mostly sand) remained in the channel. After the third slide, mudlines just reached the bank full stage suggesting that it may have taken the river slightly longer to clear its channel. It does not appear that the slide material dammed the river sufficiently that the course of the channel was diverted or the floodplain inundated.

Once failed, the slide material became extremely mobile and flowed. Several lines of field evidence suggest strongly that at least some of the fine-grained material behaved as a non-Newtonian fluid, moving until the driving force no longer exceeded its shear strength and then freezing into lobes with steep snouts reminiscent of debris flows (Figure 19). Rafts of stacked logs were pushed along by the flows and came to rest intact many meters from where they were originally stacked (Figure 20). The run out from the third slide had no debris-flow-like snouts and travelled a much shorter distance.

Additional field evidence suggests that at least some of the material in the debris apron must have been saturated when it was deposited and by analogy when it failed. Immediately after the second slide, there was spectacular evidence that dewatering had and was continuing to occur in the slide apron. Over the surface of the slide, we found dozens of fluid escape structures. These took the form of mud volcanoes with small central vents (2 to 20 cm) and large low gradient aprons (50 to 200 cm) made up primarily of fine sand (Figure 21). The vent areas of these volcanoes were easily liquifiable with agitation for over a week after the slide occurred suggesting that dewatering happened only slowly through the fine-grain material. It is possible that some Brewster River water was incorporated by the slide as it ran over the channel. The best evidence for such incorporation were splash marks we observed on the trees and a house on the upstream edge of the debris apron.

The mobility of material in the slide apron has significant implications for hazard management. The fluidization and debris-flow-like character of the debris apron allowed slide materials to move much father away from the source than we would have expected based on our experiences with other landslides in Vermont. This mobility, while it may endanger residences several hundred meters from the scarp, allowed the failed material to move away from the Brewster River channel. Less mobile slide debris might have plugged the channel, diverting the Brewster River and possibly inundating parts of Jeffersonville including the school which would likely have been moistened by waters rising behind a landslide dam.

Conclusions

The Jeffersonville slide is unusual both in its size and the fluid, mobile behavior of the slide debris. However, the stratigraphy that leads to this type of slope instability is widespread throughout Vermont and indeed throughout glaciated regions with topography sufficient to impound glacial lakes (e.g., the Puget Lowland, Dunne and Leopold, 1978). Landslide scars we have mapped on high terraces of other regional rivers such as the Huntington suggest that slides similar to this do happen elsewhere and that they are preserved both in the stratigraphic and morphologic record. Such slides are not restricted to rural, lightly developed areas. Slides of the scale of Jeffersonville have occurred repeatedly along the Winooski River in Burlington in a similar sequence of glacial-lacustrine silt overlain by sand (Figure 22) With little knowledge of this hazard, planners react retroactively rather than proactively allowing houses and businesses to be built in unstable areas and then condemning the properties after failures occur and the buildings are deemed unsafe. Better appreciation of landslide hazards in Vermont, should lead to more proactive response in terms of zoning, set back ordinances, and land use decisions.

Figure 13. (a) Portion of the Jeffersonville 7.5-minute quadrangle map showing the village of Jeffersonville and the location of Cross-section A–A'. Map width is 1.3 km. (b) Portion of the 1979 Jeffersonville orthophoto map (Sheet No. 124236 original scale 1:5000) showing an enlarged area of the topographic map. Arrow points to slide scar present in 1979 (bare area extending from the Brewster River up to the top of the terrace)from the tennis courts now buried by runout debris. North is to top in both figures.

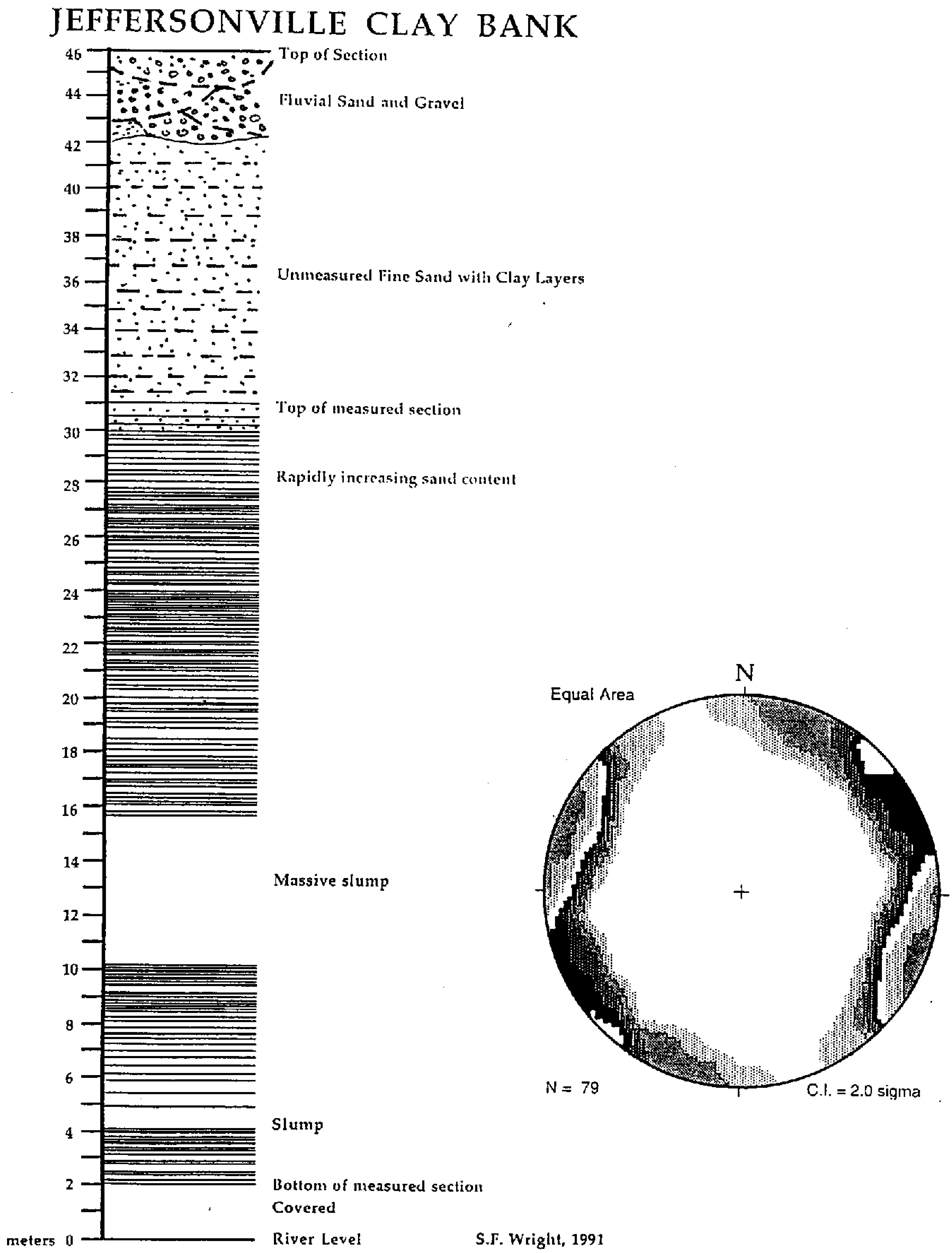


Figure 14. (a) Stratigraphic column of glaciolacustrine sediments exposed at the site of the Jeffersonville landslide. Individual clay layers marking the top of each varve are shown with a solid horizontal line. See text for detailed description. Insert shows lower hemisphere equal area stereonet plot of joints (2_ contour interval) measured in the 1 – 2 cm thick clay layers. Note that the near-vertical joints occur in two well-defined orthogonal sets. The most common strikes NNE–SSW, and the somewhat less common set strikes WNW–ESE.

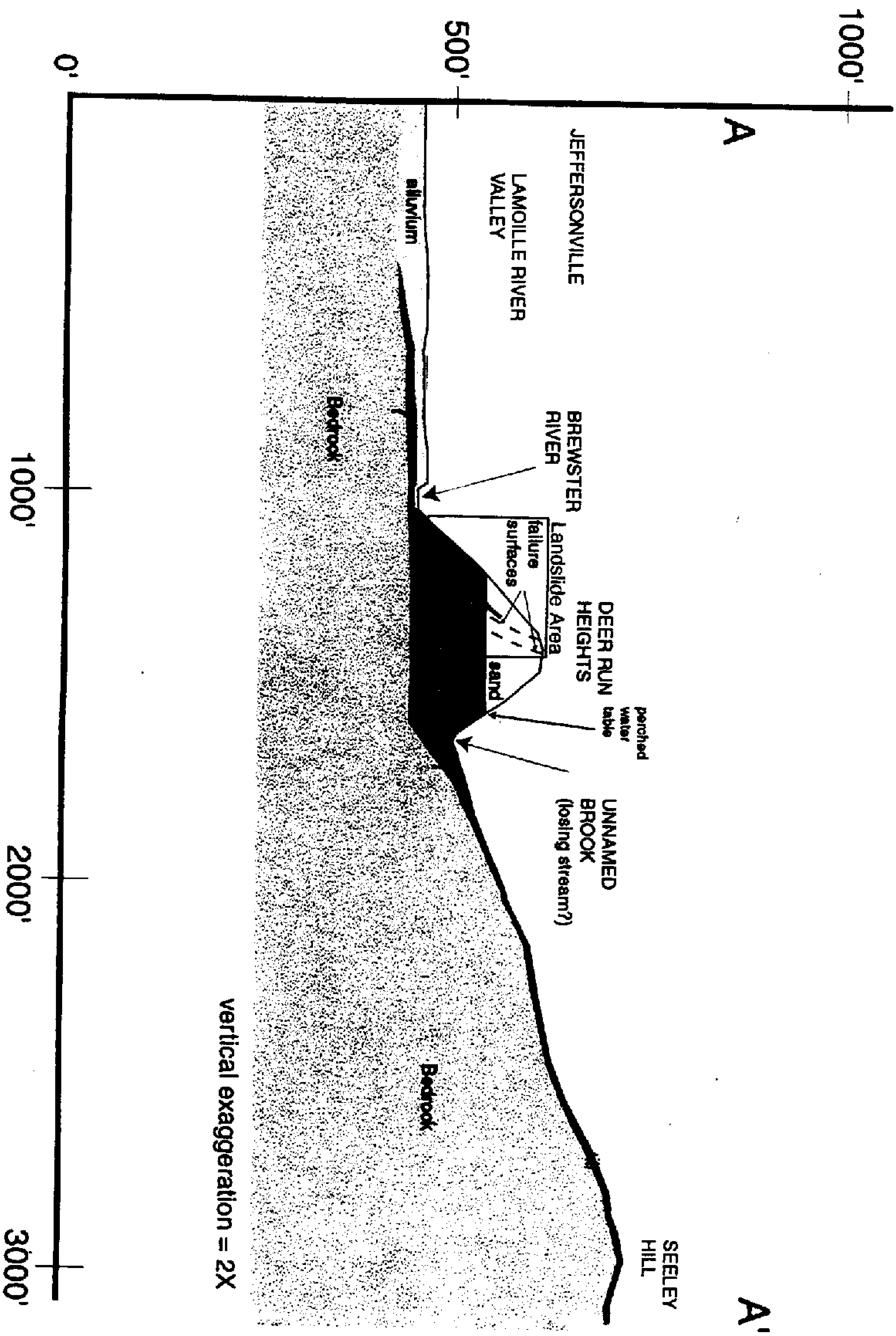


Figure 15. Idealized cross section of landscape near Jeffersonville slide. Drawn by Larry Becker, Vermont State Geologist.

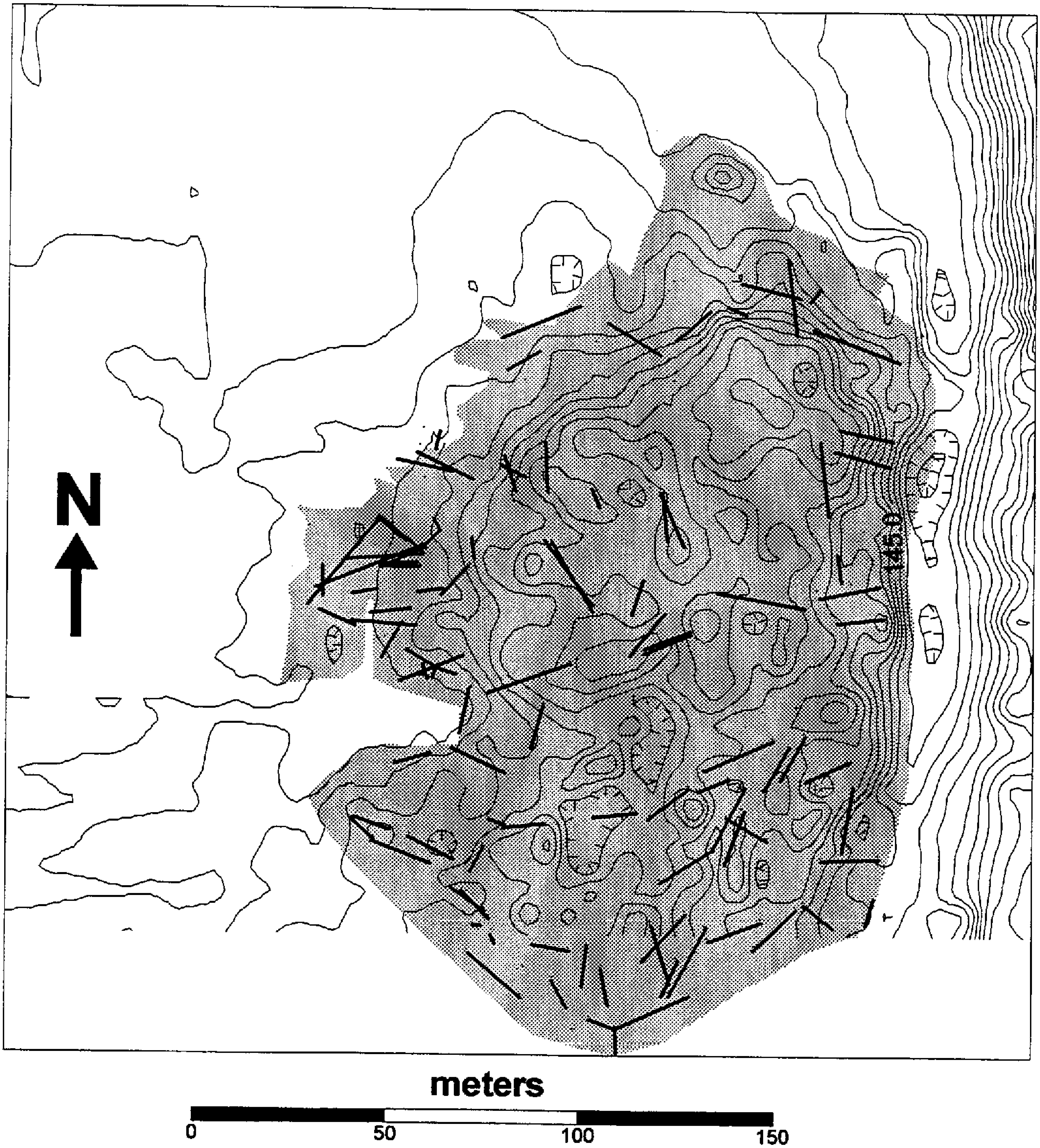


Figure 16. GPS-based map of Jeffersonville slide showing orientation of trees carried by the slide into the run out cone. Trees are represented by black lines connecting surveyed tree endpoints. Run out zone is shaded grey.

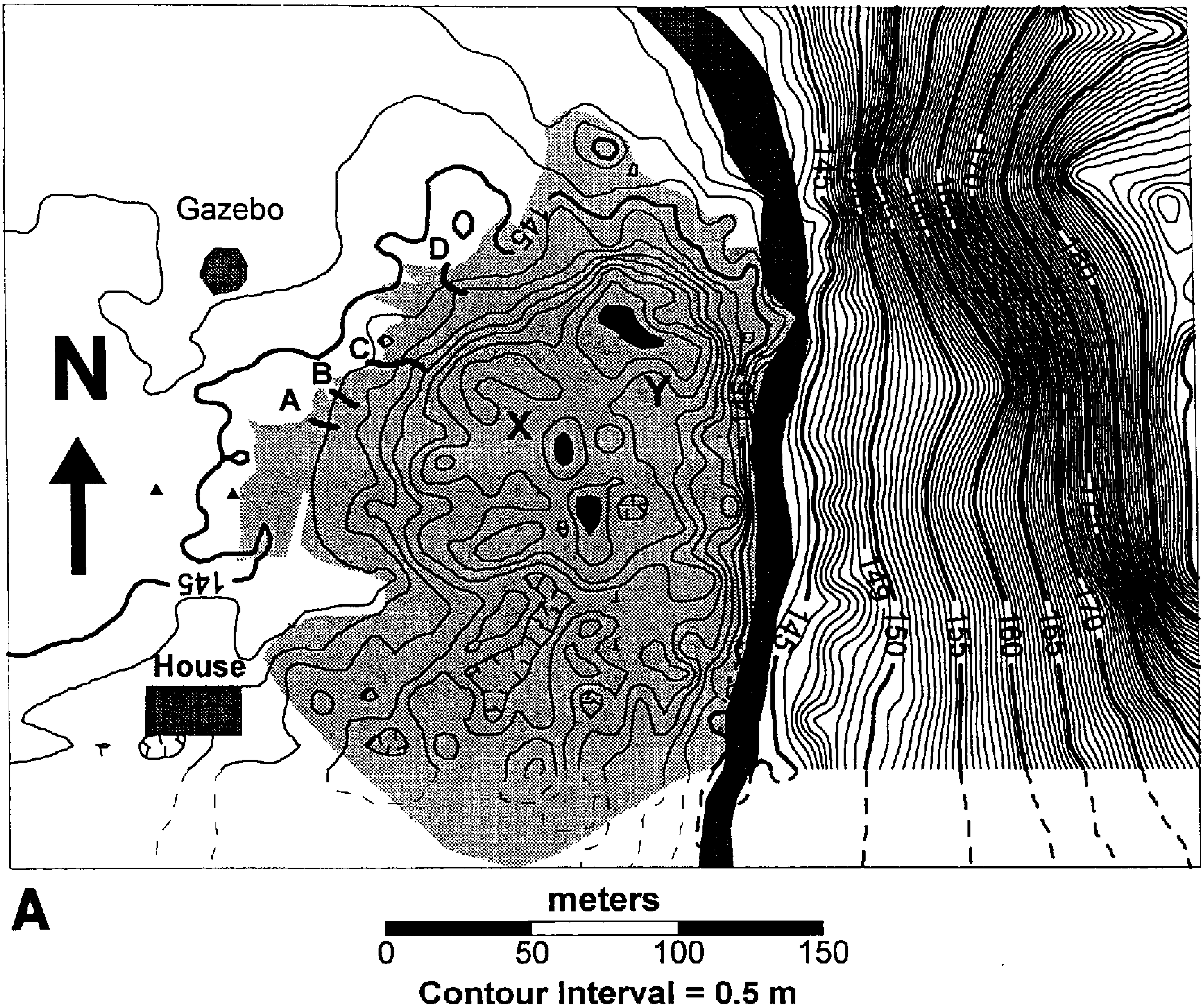
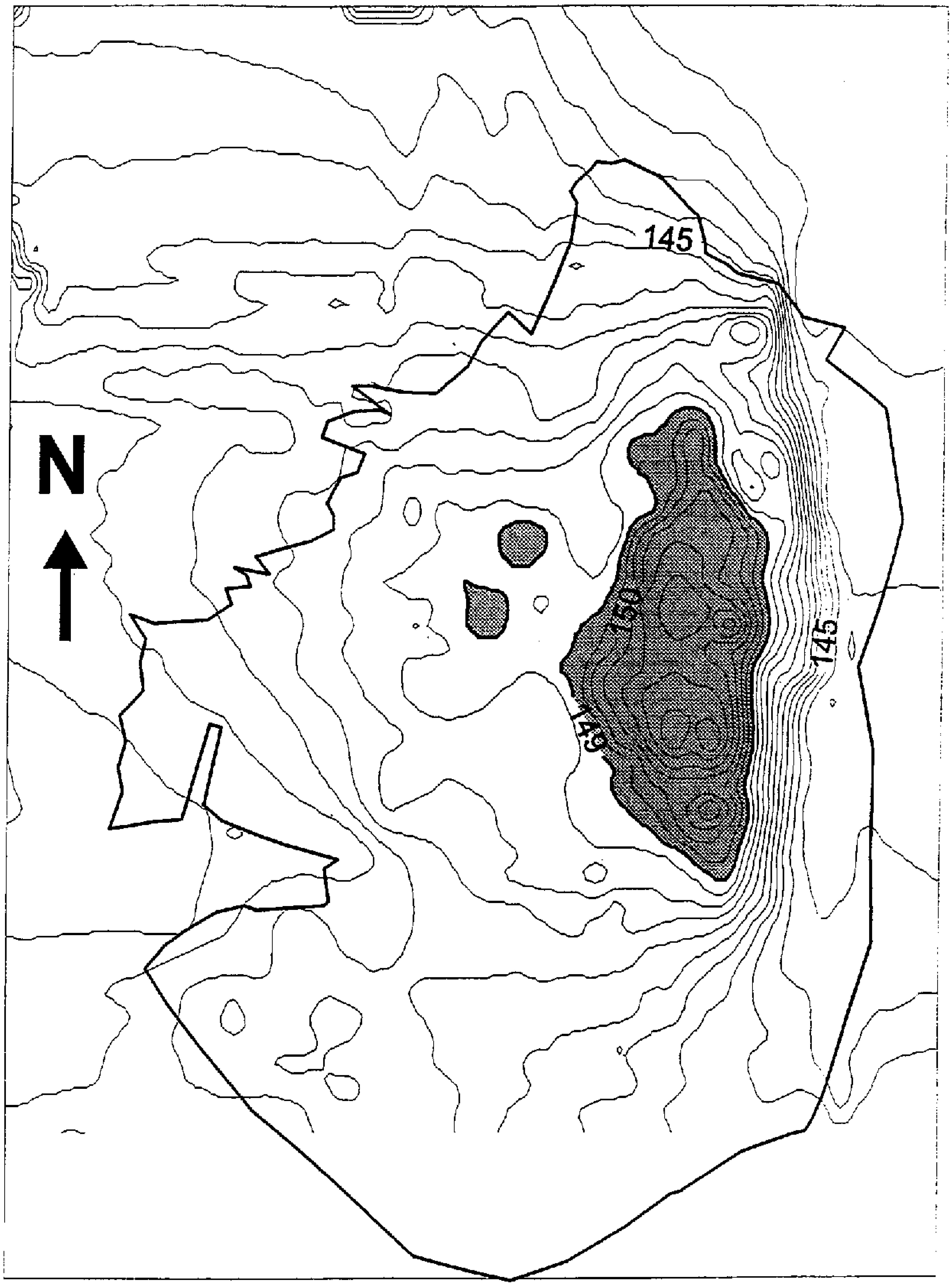


Figure 17. Topographic maps of Jeffersonville Slide. A. Map made from data collected after second slide, May 1999. Volume of slide material in run out zone is about 23,000 m^3. Debris flow snout transects across slide margins are identified by letter. Largest mud volcano indicated by "Y". Mud volcano field indicated by "X". Elevations above 149 m on run out zone are shaded black. B. Map made from data collected after third slide, July 1999. Additional volume of slide material delivered to the run out zone by this slide was about 4,200 m^3. Shaded area is hummocky and assumed to be debris from third slide, not reworked material from bulldozer. Outline of run out zone is black line. Topography is accurate (>1500 survey points) for shaded area and adjacent channel bank only.


145
150
149
145
N
B
meters
0 20 40 60
Contour Interval = 0.5 m

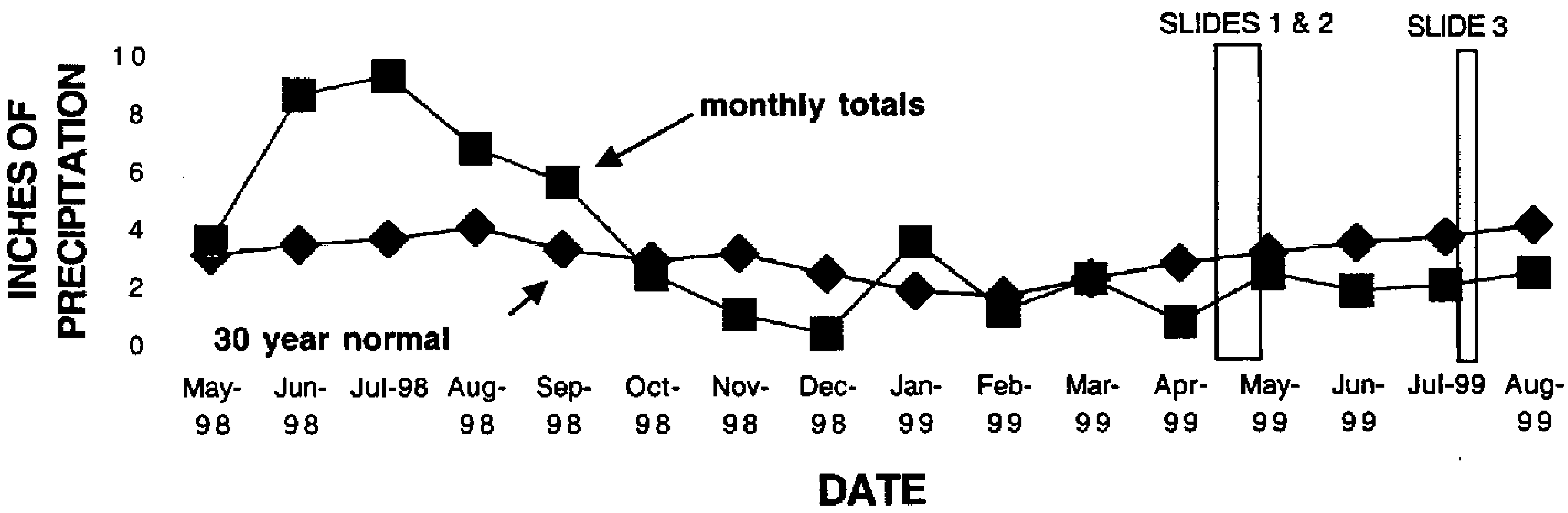


Figure 18. Monthly precipitation data for Burlington, Vermont, station BTV. Source: National Weather Service.

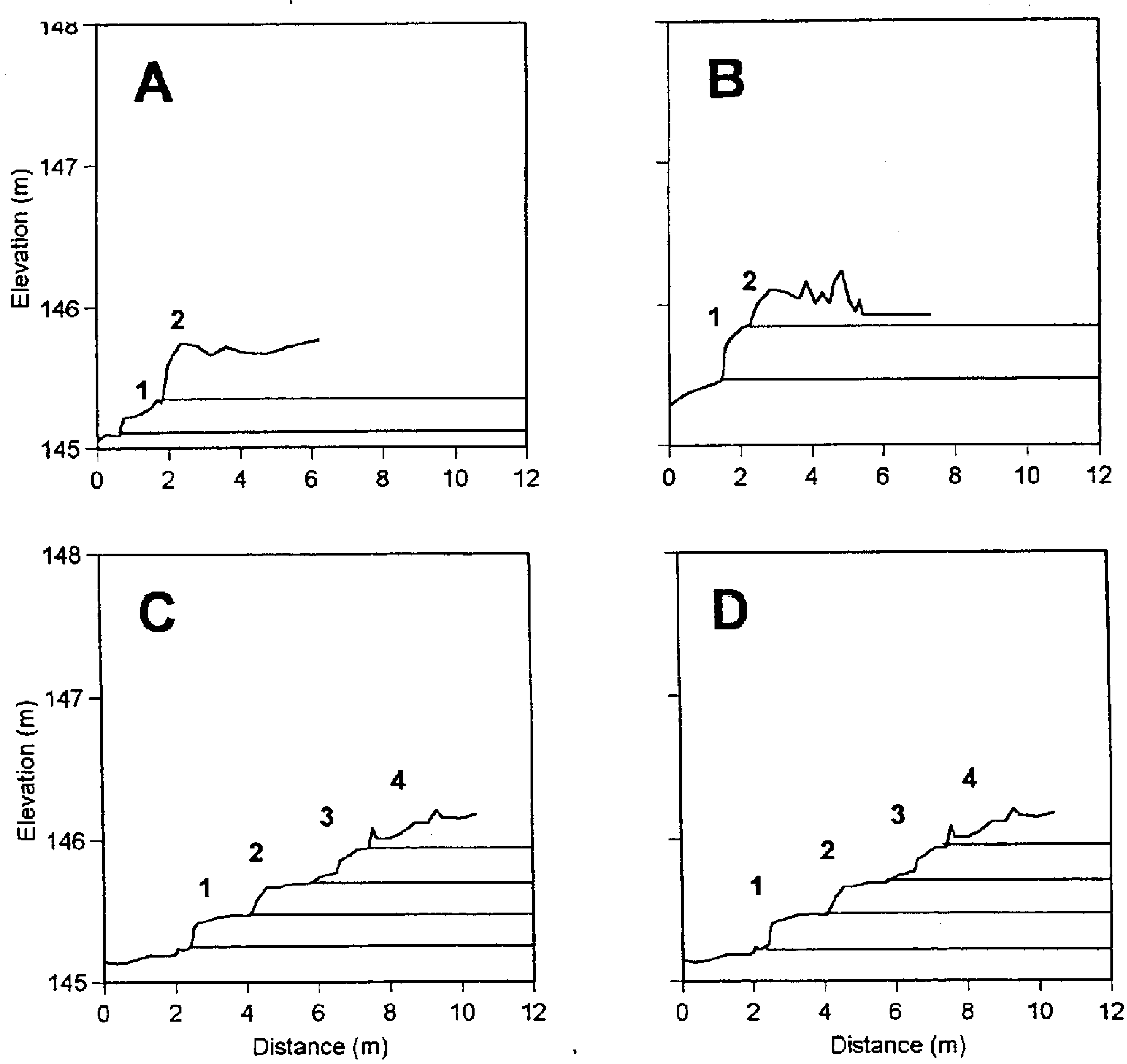


Figure 19. Cross-section of debris flow snouts at the margin of the Jeffersonville slide. Numbers indicate identifiable snouts. Snout transects locations are labeled on Figure 17A. Vertical exaggeration is 4X.

Figure 20. Photograph of log jam pushed ahead of the slide.

A

Figure 21. Photographs of fluid escape structures (mud volcanoes). A. Field of mud volcanoes ranging from 5 to 20 cm in diameter at location X on the map in figure 17A. B. Largest mud volcano observed on slide surface, at location Y on figure 17A.


B

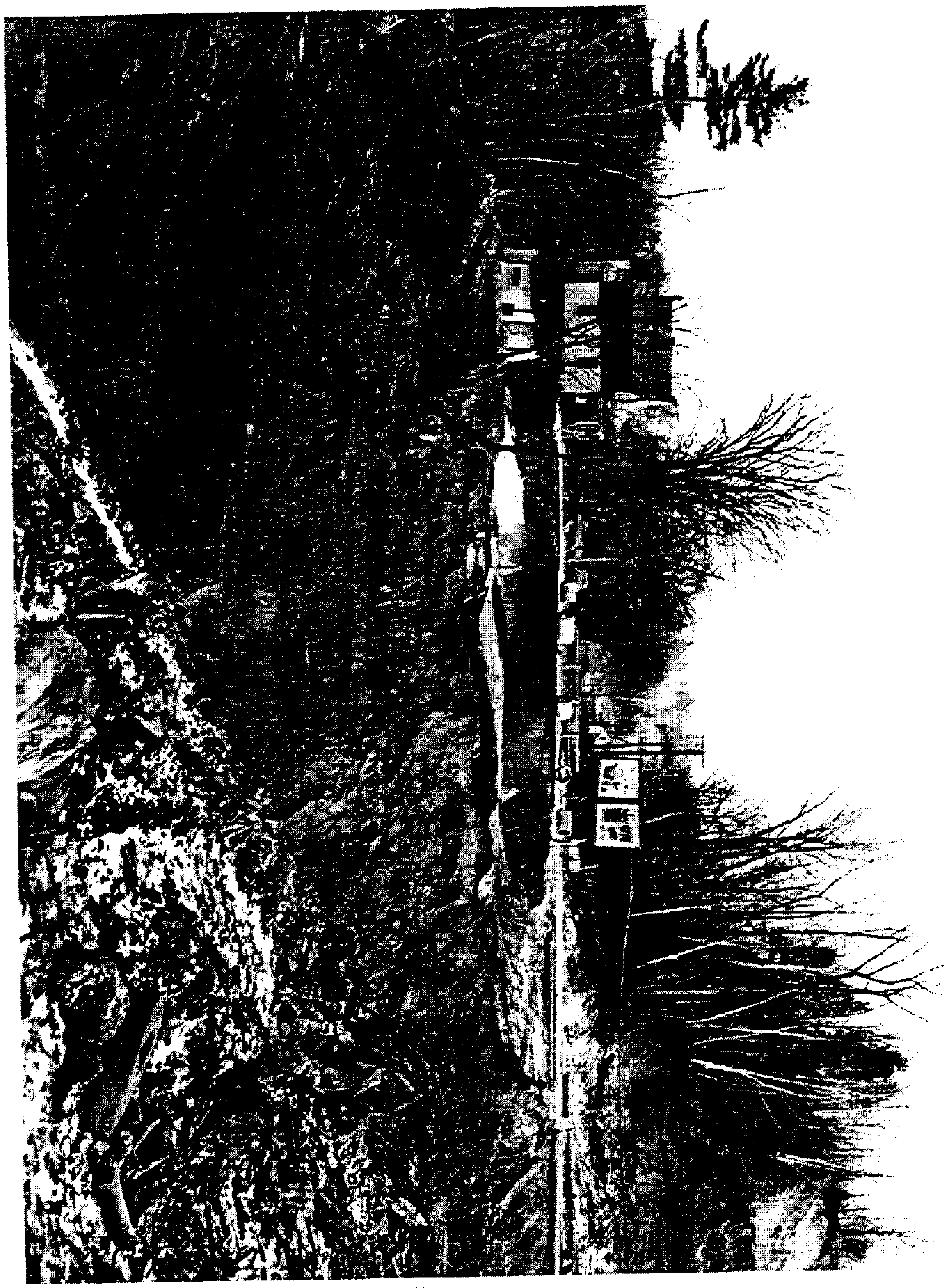


Figure 22. Photograph of 1954 landslide on Riverside Avenue in Burlington. Failure in similar materials to the Jeffersonville slide. Photograph courtesy of the University of Vermont Special Collections, Bailey Howe Library.

End of Field Trip. Return to Burlington via either Route 15, or Routes 104, 104A, and I-89.

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