INQUA 2003 Field Guide Volume

Quaternary Geology of the United States

Edited by Don J. Easterbrook

QUATERNARY GEOLOGY OF THE UNITED STATES

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Don J. Easterbrook
Dept. of Geology
Western Washington University
Bellingham, WA 98225

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Cover: Cirques and glaciated peaks in the North Cascades Washington. Photo by D.A. Rahm, courtesy of EPIC.

Contents

Foreword

Much of the landscape in the United States was shaped by climatic events during the Quaternary, especially the erosion and deposition in the Pleistocene. The wealth of Quaternary features found in the U.S. includes continental ice sheet glaciation, alpine glaciation, marine shorelines, marine deposits, faulting, tectonic uplift, effects of isostatic rebound, pluvial lakes, large scale eolian deposits, the world's largest floods from bursting of glacial lake Missoula, the world's largest volcanic eruptions, the world's largest geothermal area, and much more.

This volume contains field guides for 17 field excursions in the U.S. It consists of contributions from 97 authors across the country from Alaska and the west coast to New England, including much new, previously unpublished information. Each field guide includes specific sites with interpretations of the features to be seen and discussions of critical issues.

I would like to thank all of the authors, field trip organizers, and leaders for the thousands of hours of work that went into making this volume. I would especially like to thank Heather Sutphin at the Geological Society of America for her efforts in designing and laying out the material and for her patience in dealing with all of the details from such a large number of authors. We all hope that you will enjoy using the field guides and discover many new any interesting features of the Quaternary geology of the U.S.

Don J. Easterbrook

Quaternary and geomorphic processes and landforms along a traverse across northern New England

Authors and Leaders: **P.T. Davis,** *Dept. Natural Sciences, Bentley College, Waltham, MA 02452-4705*
P.R. Bierman, *Dept. Geology, Univ. of Vermont, Burlington, VT 05405-0122*
J. Brigham-Grette, *Dept. Geosciences, Univ. of Massachusetts, Amherst, MA 01003*
D.M. Fitzgerald, *Dept. Earth Sciences, Boston Univ., Boston, MA 02215*
B.K. Fowler, *North American Reserve, 67 Water St., Suite 207, Laconia, NH 03246-3300*
M.J. Retelle, *Dept. Geology, Bates College, Lewiston, ME 04240-6084*
J.C. Ridge, *Dept. Geology, Tufts Univ., Medford, MA 02155*
W.B. Thompson, *Maine Geological Survey, 22 State House Station, Augusta, ME 04333-0022*
T.K. Weddle, *Maine Geological Survey, 22 State House Station, Augusta, ME 04333-0022*

INTRODUCTION

This 7-day trip through northern New England begins and ends at Logan International Airport in Boston. We will examine raised glaciomarine deltas and related landforms, including moraines along the Maine coast, eskers in western Maine, landforms of alpine glaciers and continental ice sheets, along with mass-wasting deposits, in the White Mts. of New Hampshire and the Green Mts. of Vermont, and the famous varves of glacial Lake Hitchcock in the Connecticut Valley from northern New Hampshire and Vermont to central Massachusetts. We will debate the most recent chronologies derived from ^{14}C dating, varve counting, and cosmogenic exposure dating. We will also demonstrate GPR and useful field mapping techniques, and discuss various interpretations of glacial recession in New England, such as the morpho-sequence concept.

Because of the long lead-time required in the writing of this guidebook, some of the planned stops may no longer be available, so we may need to substitute some new sites. We are most grateful to landowners, too numerous to mention here, for permission to access their property to examine exposures of surficial deposits.

Some ages provided in this field trip guidebook are given in ^{14}C yrs. B.P. and others in calendar yrs. B.P.

DAY 1. DEGLACIAL AND MARINE DEPOSITS OF THE CASCO BAY LOWLAND - LOWER ANDROSCOGGIN VALLEY (Weddle and Retelle)

INTRODUCTION

The recession of the late Wisconsin, Casco Bay sublobe of the Laurentide Ice Sheet in southwestern Maine is represented by glaciomarine sediments deposited in an ice-marginal sea from a marine-based ice sheet, along with nearshore deposits associated with sea-level and postglacial isostatic rebound. Timing of late-glacial events and ice retreat is constrained by detailed mapping and stratigraphically-controlled, ^{14}C ages related to ice-marginal positions and marine-emergence deposits (Weddle et al., 1994; Weddle and Retelle, 1995, 1998). The altitudes of ice-marginal deltas and marine limit indicators, such as raised beaches deposited during ice retreat, provide a record of relative sea-level fluctuations and characterize the nature of postglacial uplift and emergence of the region (Thompson et al., 1989; Koteff et al., 1993; Barnhardt et al., 1995).

On DAY 1 (Fig. 1), we will examine deposits of late Wisconsin ice retreat and marine transgression/regression. Revised deglaciation and sea-level chronologies are based on recent detailed surficial mapping and stratigraphic studies that allow comparison with offshore records in Maine (Barnhardt et al., 1997; Schnitker et al., 2001) and onshore/offshore records of marine-based ice retreat in eastern Maine and adjacent Nova Scotia (Dorion et al., 2001; Stea et al., 2001). In addition, the Casco Bay chronology provides a link for comparison with terrestrial-based ice retreat through central New England and the White Mts. (Ridge et al., 1999, 2001; Thompson et al., 1999).

LATE WISCONSIN GLACIATION, DEGLACIATION, AND RELATIVE SEA-LEVEL

The southwestern margin of the Laurentide Ice Sheet in New England had a lobate geometry controlled by subglacial topography. The western Gulf of Maine (Fig. 1) was inundated by ice from the Charles-Merrimack lobe in southwestern Maine (whose southern extent formed the Cape Cod Bay lobe) and the South Channel lobe (Stone and Borns, 1986).

The Laurentide Ice Sheet reached its late Wisconsin maximum in the Gulf of Maine shortly after 22,000 ^{14}C yrs. B.P. (Stone and Borns, 1986; King, 1996). A regional deglaciation model proposes the existence of an ice stream in the Gulf of Maine, fed by inland ice that occupied lowland valleys (Hughes et al, 1985; Hughes, 1987). At its maximum extent, the ice stream fed a floating ice tongue over several of the deeper basins in the gulf. Based on seismic data and microfossil evidence in sediment cores, Belknap and Shipp (1991) proposed

Davis, P.T., Bierman, P.R., Brigham-Grette, J., Fitzgerald, D.M., Fowler, B.K., Retelle, M.J., Ridge, J.C., Thompson W.B., and Weddle, T.K., 2003, Quaternary and geomorphic processes and landforms along a traverse across northern New England, U.S.A.: *in* Easterbrook, D.J., ed., Quaternary Geology of the United States, INQUA 2003 Field Guide Volume, Desert Research Institute, Reno, NV, p. 365–398.

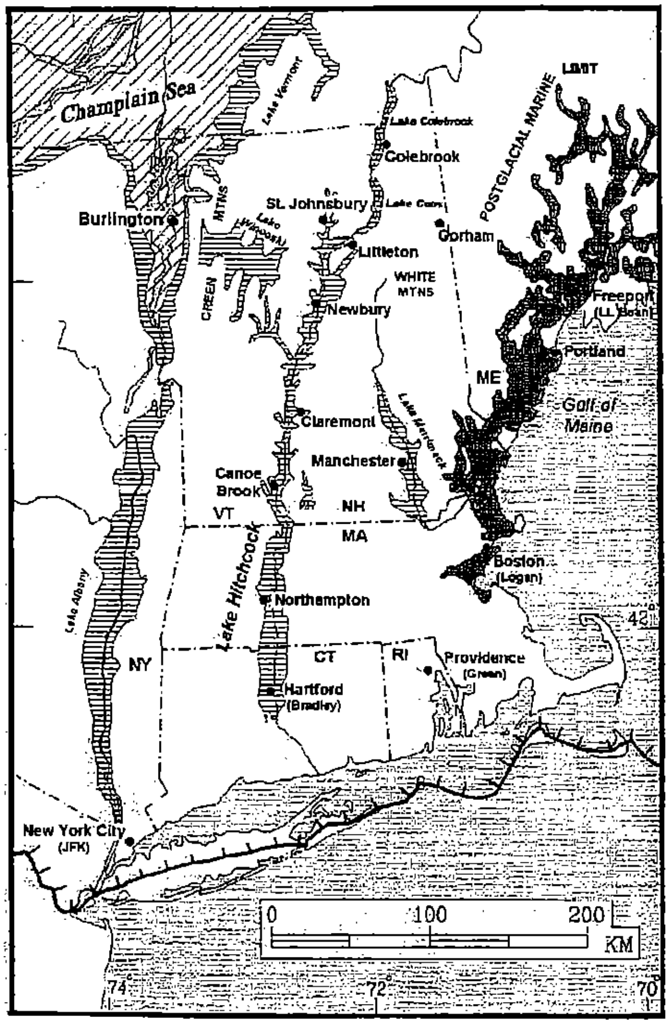

Figure 1. General location map of northern New England showing locations of larger regional airports, nightly accommodations and numbered stops for field trip, former glacial lakes (hachured), and areas that underwent postglacial marine submergence (shaded). Heavy hachured line shows LGM limit of Laurentide Ice Sheet (modified from Ridge, 2001).

that during initial deglaciation of the outer continental shelf, the ice shelf was pinned in several places on bedrock topographic highs in the Gulf of Maine.

Schnitker (1986) estimates that initial deglaciation in Wilkinson Basin in the central Gulf of Maine occurred prior to 17,600 ^{14}C yrs. B.P. To the northeast, Gipp and Piper (1989) and King (1996) place the deglaciation of the central Emerald Basin south of Nova Scotia between 17,700 and 18,000 ^{14}C yrs. B.P., although sediments interpreted as ice-proximal are found on the margin of the basin as late as 15,000 ^{14}C yrs. B.P.

Hughes et al. (1985) and Hughes (1987) proposed that a calving bay in the Gulf of Maine replaced ice streams and floating ice shelves by 14,000 ^{14}C yrs. B.P. On the inner continental shelf, Belknap and Shipp (1991) and Bacchus (1993) describe a glaciomarine unit that drapes over bedrock and is overlain by a horizontally-bedded glaciomarine mud. They interpret this stratigraphy to represent the shift from a floating ice margin to a calving embayment during deglaciation. As ice retreat progressed, calving along the floating margin eventually gave way to a grounded ice margin with calving along a tidewater cliff, similar to the conditions reported from glaciomarine tidewater settings in southeastern Alaska (Molnia, 1983; Phillips et al., 1991; Powell 1991).

Alternatively, Oldale (1989) suggests that ice was grounded at its maximum position in the western gulf and that the glaciomarine drift was deposited from point sources at the grounded tidewater terminus of the icesheet. Rapid deglaciation of the South Channel and Cape Cod Bay lobes may have by a calving front over Wilkinson Basin in the western Gulf of Maine (Stone and Peper, 1982).

Regardless of the mode of ice retreat on the continental shelf, numerous studies of glacial deposits on land in Maine and adjacent regions have demonstrated that moraines and grounding line fans (cf. Powell, 1991) are interbedded with glaciomarine sediments, inferring that deglaciation of the coastal lowland by a grounded tidewater glacier was contemporaneous with marine submergence of the region (Borns and Hagar, 1965; Stuiver and Borns, 1975; Thompson, 1982; Smith, 1985, Retelle and Bither, 1989, Ashley et al., 1991). Lowell and Borns (1994) have suggested that the effectiveness of a calving embayment in the major valleys across Maine was likely impeded by topography and shallow water depths of the transgressing sea, and that deglaciation may have been more influenced by climate at that time. By 14,000 to 12,000 ^{14}C yrs. B.P., the ice waned, and smaller sublobes of the retreating icesheet occupied the lowlands. We use the name Casco Bay sublobe for the lobate ice mass that occupied the Casco Bay Lowland.

Shoreline features, such as deltas and beaches, which define the local marine limit increase in elevation from the coast to the interior of Maine in response to the former increased isostatic load of late Wisconsin ice (Thompson et al., 1989). In the field area, the marine limit increases from about 75 m along the outer coast to over 100 m in the upper Androscoggin Valley (Thompson et al., 1989). The marine limit was established synchronously with ice retreat, so the oldest marine sediments transgressed the lowland from 14,000 to 13,000 ^{14}C yrs. B.P.

The marine transgression was followed quickly by regression of the synglacial sea as relative sea level dropped due to rapid glacio-isostatic uplift (Bloom, 1960, 1963; Borns and Hagar, 1965; Schnitker, 1974; Stuiver and Borns, 1975; Belknap et al., 1987; Crossen, 1984, 1991; Kelley et al., 1992; Barnhardt et al., 1995). Koteff et al. (1993) estimated that emergence in coastal central New England began sometime between 14,000 and 13,300 ^{14}C yrs. B.P. Smith (1985). Smith and Hunter (1989) estimate maximum marine inundation of the coastal lowland by 12,600 ^{14}C yrs. B.P. and emergence of the present landmass of southwestern Maine by 11,500 ^{14}C yrs. B.P. The postglacial low stand of the sea extended onto the inner continental shelf to a depth of -55 m by 10,500 ^{14}C yrs. B.P. (Barnhardt et al., 1997).

STOP 1-1. BRUNSWICK SAND PLAIN, BRUNSWICK, MAINE

At this stop, we will examine a section of nearshore bioturbated sediments overlain by fluvial deposits. These deposits are the upper part of the Brunswick sand plain, a regressive coastal delta (Retelle and Weddle, 2001; Weddle 2001).

The surficial geology in the Brunswick quadrangle records the advance and retreat of the Laurentide Ice Sheet, and subsequent Holocene events and deposits. The Brunswick sand plain is the most prominent deposit associated with this record (Fig. 2), with ^{14}C ages on marine fossils dating the formation of the sand plain to a time of worldwide rapid rise of sea level.

As ice sheets retreated at the end of the late Wisconsin and the Laurentide ice margin reached the present-day coast of Maine, the synglacial sea flooded the Gulf of Maine, making contact with the ice front. This transgression was in response to isostatic downwarping. As the ice melted, the depressed crust did not respond immediately to the release of weight from the ice, and as a result the sea flooded inland, up river valleys and across lowlands. In the Brunswick quadrangle, the marine limit is ~84 m above present sea level. During ice retreat and sea-level rise, meltwater discharging from the ice deposited sediment in the sea. Later, isostatic emergence of the land resulted in nearshore deposits associated with dropping, relative sea level.

The elevation of the sand plain is as much as 53 m below the highest elevation of older marine deposits, indicating that the sand plain is younger than the marine sediments. Exposures in the surface of the plain show trough-cross-bedded sand, typical of braided streams. Boring logs and geophysical data record the transition of late-glacial, isostatic emergence from marine to nearshore conditions and deposition of the Brunswick sand plain. Surficial sand of the plain overlies a sandy-silt zone that includes discrete sand units that dip gently eastward. Beneath the sandy deposits, thick glaciomarine mud overlies sand and till lying on bedrock.

Marine fossils just above the highest surface of the sand plain yielded an age of 12,800 ^{14}C yrs. B.P. Marine fossils from an offshore, seismically-identified sequence provide the youngest ages for the sand plain (12,200 ^{14}C yrs. B.P.; Oakley, 2001). The seismically identified units are likely the distal equivalent of the sand plain.

Marine fossil ^{14}C ages must be corrected for the ^{14}C marine reservoir effect (lag time for atmospheric ^{14}C to mix with sea water). For the Pleistocene Gulf of Maine, the correction is estimated at ~-600 to -700 years (Dorion et al., 2001). Thus, using 12,200 and 11,600 ^{14}C yrs. B.P. as corrected bracketing ages for the time of formation of the Brunswick sand plain, the time overlaps the period of worldwide sea level during meltwater pulse-1A, which occurred between 12,600 and 11,700 ^{14}C yrs. B.P. (Bard et al., 1990; Adkins et al., 1998). The data support the interpretation that the sand plain is a coastal delta formed as emergence continued, but during a period of relative sea-level stability, when the rate of local isostatic emergence kept pace with eustatic rise of sea level.

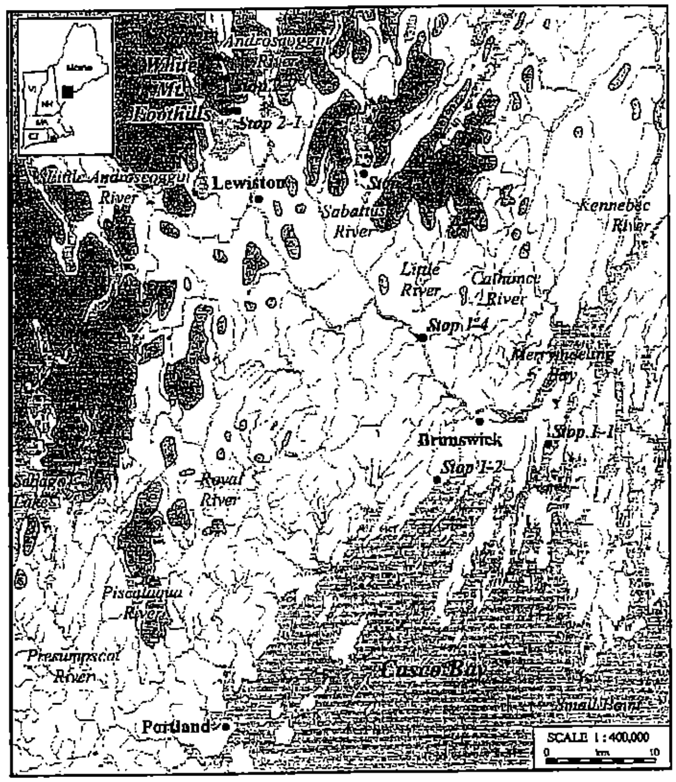

Figure 2. Field trip area for Days 1 and 2 (Stops 1.1 to 1.4 and 2-1 to 2-3). Gray shaded area is land above 100 m elevation above sea level (approximate elevation of local marine limit).

STOP 1-2. BUNGANUC BLUFF, BRUNSWICK, MAINE

This stop can be very muddy! Several large slump exposures along the coast reveal a section of glacial, glaciomarine, and nearshore deposits that represent changing environments in southwestern Maine (Fig. 2). Subglacial and ice-marginal sediments are present at the west end of the bluff, overlain by a thick, muddy, medial-to-distal glaciomarine fan or submarine plain in the central and eastern end of the section. In turn, nearshore deposits and an eolian cap overlie these. Belknap et al. (1987) and Belknap and Shipp (1991) described the section as an example of glaciomarine mud deposited in part under an ice shelf or at least near the ice-grounding line. Graded, rhythmically-bedded, fine sand. silt, and clay exposed here and identified elsewhere in subsurface boring data represent distal deposition with cyclical sorting of fine sediment. The lower deposits are most likely glaciomarine sediments, whereas the upper deposits are more likely marine and nearshore sediments. One question to consider at this stop is the influence of cyclical process on discharge, such as daily, seasonal, and tidal processes as described by Cowan and Powell (1990) and Phillips et al. (1991). A better understanding of the depositional environment at this exposure would allow comparisons with modern glaciomarine environments.

At the west end of this site, a very-poorly-exposed section consists of a thin veneer of sandy till (represented by numerous boulders at the base), massive sand, and glaciomarine mud, which overlie striated (170°) bedrock (Weddle, 1994). Eastward, near the top of the section is 2-3 m of medium-to-fine-grained sand, which contains rip-up mud clasts. The sand is found in erosional channels cut into underlying syn-depositionally-deformed, stratified sand, silt, and clay. In gradational contact with the upper sandy portion of the bluff, lower muddy sediments are best exposed in slumps at the eastern end of the section. These lower deposits consist of thin, well-stratified, laterally-continuous, silt and clay beds with silty, fine-sand laminations, grading upward into more massive mud. Dropstones are rare. At the east end of the section, striated (195° to 201°) bedrock is exposed at the waterline. Southwest-oriented striae are found along many of the peninsulas and islands in this part of Casco Bay and reflect the strong topographic control on late Wisconsin ice by the northeast trend of bedrock structure in the area. Seismic surveys show that depth to bedrock in the bay troughs east of the bluff is over 300 feet in places, the deep northeast-trending troughs preferentially controlling ice flow.

LUNCH STOP: Bradbury Mt. State Park, Pownal, Maine

STOP 1-3. DELTAS, FANS, AND MORAINES, SABATTUS, MAINE

Along the east side of Sabattus Lake (Fig. 2), we will examine fine examples of glaciomarine fans and deltas, and minor moraines (cf Ashley et al., 1991; Powell, 1991). The collective assemblage forms a beaded morphology (Sollid and Carlson, 1984; Caldwell et al., 1985) between two larger Gilbert-type deltas along the east side of Sabattus Pond. The two deltas and a series of smaller, elongate, lobe-shaped fans represent distinct ice-marginal positions where submarine outwash emanated from the ice-tunnel system during retreat of the ice. As many as four or five small moraines flank each of the beads along the east side of the valley. The moraine ridges have a general east-west alignment, are convex down-valley, and typically cluster in the valley, pinned to higher hillocks on the adjacent valley upland to the east. As many as 40 minor moraine ridges, with an average spacing of 50 to 100 m, are mapped in the 2-km-long valley east of Sabattus Pond (Bernotavicz, 1994). These moraines, which usually occur in clusters, are 3 to 10 m high, 30 to 50 m wide, and hundreds of meters long, but their internal composition is unknown.

The two deltas in the valley have different morphologies but share some noteworthy characteristics. The relatively flat surface elevation of each delta is about 100 m. Both have an elongate northern end, most prominent on the Marr Point delta. A ridge, traceable along Pleasant Hill Road, enters the north end of the Pleasant Hill delta just east of Round Pond (a kettle). These features represent the position of an ice-tunnel at or near the ice margin where sediment emanated from the tunnel to build the deposits. The Pleasant Hill delta is kidney-shaped, more lobate than the Marr Point delta, which is more wedge-shaped. The Pleasant Hill delta likely represents laterally migrating and coalescing delta or fan lobes along the ice margin. The Marr Point delta was apparently deposited with little or no lateral migration of a tunnel feeder source. Bernotavicz (1994) noted a change in orientation of end moraines near the Marr Point delta front and suggested that the shift from a roughly east-west trend (parallel to the ice margin) to a northwest-southeast trend reflects opening and enlargement of the ice tunnel, allowing lateral expansion of the Marr Point delta. During deglaciation, as the ice margin retreated in the Sabattus Pond lowland, the tunnel was pinned to the east at Sabattus Mt..

STOP 1-4. FAN DEPOSIT, MAINE DOT PIT, TOPSHAM, MAINE

The purpose of this stop is to see a Pleistocene, nearshore deposit associated with regression of the sea after uplift was underway and to discuss ^{14}C chronology in southwestern coastal Maine (Fig. 2). The deposit overlies glaciomarine mud, which rests on distal fan deposits of interbedded silt and fine sand. The larger exposure in the core of the fan immediately adjacent to the north shows the coarse nature of the proximal fan deposits associated with the ice tunnel from which the fan deposits originated. The flanks of the fan are mantled by massive glaciomarine mud from which the pit operator claims fossil shells were found. Bloom (1960) reported a varied assemblage of fossil shells from this pit, including *Hiatella arctica, Macoma calcarea, Musculus substriata, Mya arenaria, Mytilus edulis, Nuculana jacksoni, Serripes groenlandicus, Natica clausa, Neptunea decemcostata*, and *Balanus*, representing mixed intertidal and deep-water affinities. Molds of pelecypods have been found in the mud under the nearshore deposit, but fossils have not been observed recently in the pit. Retelle and Weddle (2001) report an age of 13,240 ± 190 ^{14}C yrs. B.P. (GX1632) on *Mesodesma arctatum* from the adjacent gravel pit to the north. In context with other earlier reported ages in the region (Attig, 1975; Stuiver and Borns, 1975), these ages bracket the marine regression between about 13,300 to 11,500 ^{14}C yrs. B.P.

DAY 2. DEGLACIATION FROM THE MARINE LIMIT TO THE WHITE MOUNTAINS, WESTERN MAINE AND EASTERN NEW HAMPSHIRE (Thompson)

On Day 2, we will start near Lewiston, Maine, close to the inland limit of late-glacial marine submergence and proceed northwestward across the foothills of the White Mts. (Fig. 2). This part of the trip will examine glaciomarine deltas, eskers, eolian deposits, and a group of moraines on the Maine-New Hampshire border.

Seaward from the marine limit, esker systems are punctuated by deltas and submarine fans where sediments washed into the sea at the mouths of ice tunnels. These deposits resemble the "DeGeer eskers" in Scandinavia. The contacts between topset and foreset beds in the marine deltas record sea level during deglaciation. Inland from the zone of marine submergence, many esker systems consist of long, variably segmented ridges that are locally associated with glacial lake deposits.

STOP 2-1. AUBURN PLAINS DELTA, AUBURN, MAINE

The gravel pit at STOP 2-1 (Fig. 2) is located in one of the many ice-contact glaciomarine deltas that formed as the Laurentide Ice Sheet receded from Maine's coastal lowland (Thompson et al., 1989). The Auburn Plains delta is associated with a segmented esker system that extends from Lewiston north to Sumner (Thompson and Borns, 1985).

Glaciomarine deltas in Maine are the Gilbert type, consisting of gravelly topset beds (fluvial delta-plain deposits) overlying sandy to gravelly foreset beds that accumulated on the prograding delta front. The contact between the topset and foreset beds marks the sea level to which the deltas were graded. Exposures at STOP 2-1 have shown up to ~2 m of poorly-sorted topset gravel with local cut-and-fill fluvial structures indicating generally southward flow. The foresets consist of well-stratified sand and gravelly sand beds that dip between northeast and south-southwest. The topset/foreset contact, surveyed at three locations in the pit, ranges in elevation from 106.4 to 107.0 m (349-351 ft). Coarse gravel, deep in the deposit as reported by the pit operator, may be a buried ice-tunnel facies.

Thompson et al. (1989) surveyed the elevations of topset/foreset contacts throughout southern Maine to determine the gradient of isostatic crustal tilt. Thompson and Koteff (2002) refined these elevation data and found that the deltas north of Sanford lie on a plane that slopes SSE at 0.50 m/km. This is a minimum value for tilt because the deltas were deposited diachronously as the ice withdrew and relative sea level fell. Fig. 3 shows the elevation profile of both the unmodified high-stand deltas (including Auburn Plains) and those that were eroded during marine regression. The lowest erosion surfaces (Fig. 3) define a stillstand position of relative sea level, which has been subsequently tilted with a gradient similar to that of the marine-limit plane.

STOP 2-2. WOOD STREET DELTA, TURNER, MAINE

The Wood Street pit (Fig. 2) is located in the same cluster of marine deltas as STOP 2-1. Contours on the Lake Auburn West quadrangle suggest that the top of the Wood Street delta has an elevation similar to Auburn Plains. This locality is in a narrow delta splay that extends westward from the adjacent esker system. The foresets consist of well-stratified sand and pebbly sand beds dipping between west-southwest and northwest. In late 2002, the upper part of the pit face showed several meters of foreset beds (10-100 cm) with small-scale cross bedding. At the base of this sequence, convex-upward beds formed a climbing dune. These structures, as well as high-energy current ripples elsewhere in the delta foresets, indicate traction currents that are unusual in the glaciomarine deltas (where flow separation occurred as fresh water spread over the denser underlying sea water). They suggest estuarine conditions with a high input of glacial meltwater in the vicinity of the ice tunnel mouth. The delta foresets in the northern part of the pit are locally overlain by ~2 m of thin planar sand beds, which are probably eolian.

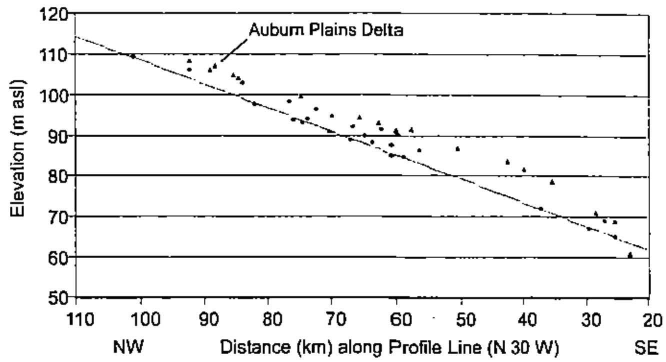

Figure 3. Glaciomarine delta elevations in southwestern Maine. Triangles indicate topset/foreset contacts of marine-limit (highstand) deltas. Diamonds show lower relative sea levels (RSL) marked by fluvial downcutting and shoreline erosion of delta surfaces. Linear regression line (r^2 = 0.98) indicates a strandline formed during stillstand of RSL (see Thompson and Koteff, 2002).

STOP 2-3. ESKER, TURNER, MAINE

The Turner town gravel pit will be visited if fresh exposures are present at the time of the field trip (Fig. 2). The working face is located in the north end of the long pit that follows the esker. Exposures reveal an east-west cross section through the esker ridge, which is about 12 m high and reaches an elevation of about 110 m (360 ft, essentially the local marine limit). Material in the pit face appears to be mostly pebble-boulder gravel, but a better exposure would likely show that the ridge consists of a coarser tunnel facies overlain by fan deposits, as is common in other eskers in the glaciomarine environment.

LUNCH STOP: Snow Falls Gorge, Little Androscoggin River, West Paris, Maine

STOP 2-4. ESKER, BETHEL, MAINE

The Hilltop pit is located in one of the major esker systems in Maine, named the Portland system by Stone (1899). This system extends discontinuously from the Quebec border southward through several large lake basins, then through a gap in the high mountains of western Maine, across the foothills, and down the Little Androscoggin Valley to New Gloucester (Thompson and Borns, 1985). A series of glaciomarine deltas formed in association with the esker as the ice margin retreated from New Gloucester to the imarine limit at South Paris (Fig. 2).

Much has been speculated about the origin of Maine's eskers and the glacial drainage networks in which they formed. Several investigators have proposed that the ice tunnels existed continuously along their entire lengths, rather than developing in shorter segments as the ice margin receded. Arguments in favor of simultaneous, full-length tunnels include (1) the well-integrated, dendritic patterns of the esker systems over their entire lengths, (2) their tendency to cut across valleys and climb over hills in response to the gradient of the ice surface, and (3) the great volumes of sediment in marine deltas at the ends of some of the systems (Shreve, 1985; Ashley et al., 1991; Boothroyd, 1995; Hooke, 2000). However, some authors propose that the deposits comprising the eskers formed near the ice margin in a south-to-north progression as the tunnel mouths receded (e.g. Ashley et al., 1991; Weddle et al., 1994). Sediment probably was delivered to the tunnels by convergent flow of adjacent glacial ice toward the tunnel axes, though field evidence of this process remains poorly documented in Maine.

The esker seen here is a narrow and nearly continuous ridge of sand and gravel along Route 232. It leaves the Androscoggin Valley and ascends the gently north-sloping valley of Barkers Brook. Stone (1899) showed a view of the esker from a nearby hill (Fig. 4). A parallel esker occurs near here on the west side of the Barkers Brook valley, and also a short distance south on Route 232. These sections of the system are examples of what have been called "multiple-crested eskers" (Shreve, 1985) or "esker nets" (Hooke, 2000). Shreve noted that they occur "in gently ascending reaches of esker paths," which is the case here, while Hooke relates esker nets to "places where the ice is so thin that tunnel closure rates cannot keep up with melt rates...".

STOP 2-5. SAND DUNE, BETHEL, MAINE

This stop is on East Bethel Road, on the south side of the Androscoggin Valley. A small barrow pit at the base of the hillslope exposes ~5 m of well-stratified sand with thin planar beds. This deposit is interpreted as eolian on the basis of its morphology, good sorting, lack of stones, and absence of sedimentary structures characteristic of waterlaid deposits.

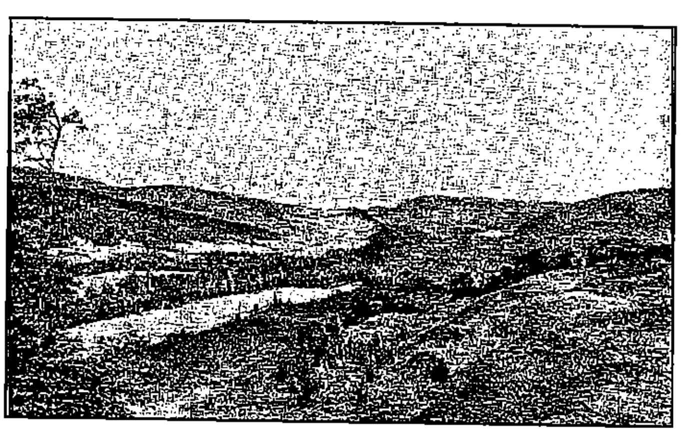

Figure 4. Illustration from Stone (1899), looking south along the esker in Barkers Brook valley from viewpoint near STOP 2-4.

Dune deposits are common on the downwind (south and east) sides of major valleys in southwestern Maine. The dunes vary in age from late-glacial to those that have been reactivated in historic time. Dunes may be parabolic or longitudinal ridges, but commonly are irregular mounds and sheets that range in thickness from 1 to 10 m or more. Till-stone ventifacts are found at the contacts of the eolian sand with underlying till or gravel. In some places, pebbles and cobbles are found within the dunes, but how they got there is uncertain.

STOP 2-6. ANDROSCOGGIN MORAINE COMPLEX, SHELBURNE, N.H.

The Androscoggin moraine complex is located on the Maine -New Hampshire border in the Androscoggin Valley (Figs. 5, 6). Stone (1880, 1899) first recognized these moraines and striations indicating local eastward ice flow. Thompson and Fowler (1984) discovered additional moraine segments on both sides of the valley. The ridges are sharp-crested, very bouldery, and up to 30 m high. The latter investigators dug backhoe test pits at the five locations (Fig. 5). Four pits east of Stock Farm Mt. showed glacial diamictons containing lenses of sand, silt, and gravel. Pit 5, near the valley axis in the lowest part of the moraine complex, revealed a waterlaid diamicton with interbedded silt, fine sand, and flowtill(?).

Stone counts from the test pits indicated that local rock types are predominant. However, basaltic and rhyolitic dike rocks totaled 16 percent on the north side of the valley (Pit 5), while samples from the south side contained only 0-2 percent of these lithologies. Bedrock mapping by Billings and Fowler-Billings (1975) shows dike swarms on the north side of the Androscoggin Valley at Gorham and farther north in the Berlin area (Fig. 6). Thompson and Fowler (1989) concluded that the northern side of the Androscoggin ice tongue in the Shelburne area would have incorporated more material from these dikes, thus explaining the variation in stone counts.

The age of the moraine complex has not been determined. Borns et al. (2003) suggest an age of about 12,800 ^{14}C yrs. B.P. Whether the moraines represent a readvance down the valley or just a stillstand during progressive ice retreat is uncertain. The moraines may have formed in response to funneling of ice flow along the valley after the late Wisconsin icesheet thinned over the crest of the adjacent northeast-trending mountain chain.

DAY 3. GLACIAL GEOLOGY OF MT. WASHINGTON AND SURROUNDING "NOTCHES," NORTHERN WHITES MOUNTAINS (Davis and Fowler)

Glacial geology of the Presidential Range

The name Goldthwait is synonymous with the glacial history of the Presidential Range (Figs. 6, 7). J.W. Goldthwait (1913, 1916a) was the first to carry out an extensive study of glaciation in the range, where he reached three major conclusions: (1) the

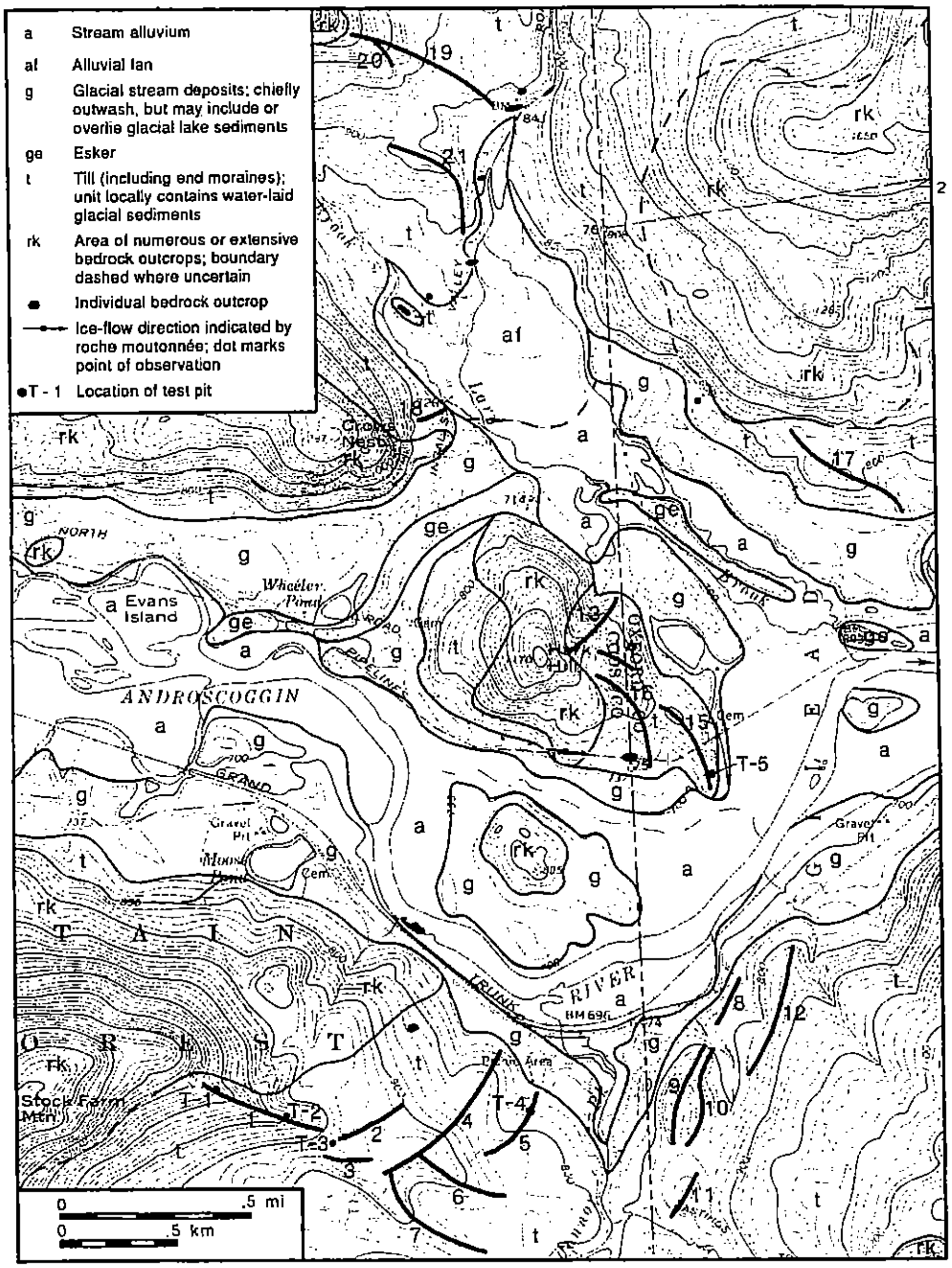

Figure 5. The Androscoggin moraine system (from Thompson and Fowler, 1989).

uplands above the cirques were eroded by both fluvial and glacial activity, (2) the cirques were carved by alpine glaciers, as opposed to continental ice, stream erosion, or frost action, and (3) continental glaciation followed the last cirque glacier activity on the range (Davis, 1999). His evidence that cirque glaciers were not active following continental glaciation included: (1) the lack of looped end moraines on cirque floors, (2) till of a northern provenance on cirque floors, and (3) asymmetric cirque cross-valley profiles. Goldthwait (1913) did not support the concept that local glaciers extended far down valleys from an icecap centered on the Presidential Range, as proposed by Packard (1867), Vose (1868), Agassiz (1870), and Hitchcock (1878).

Over the next two decades, two workers disputed the conclusions of J.W. Goldthwait concerning the timing of continental and cirque glaciation in the Presidential Range. Johnson (1933) suggested the lack of end moraines in cirques is not sufficient evidence to conclude that continental ice post-dated cirque glacier activity in the Presidential Range, as he noted other alpine areas in the world that have never undergone continental glaciation but have cirques that lack moraines. Antevs (1932) sided with Johnson, concluding that late Wisconsin cirque glaciers existed in the Presidential Range; however, neither author provided a convincing explanation for the till of northern provenance on the cirque floors.

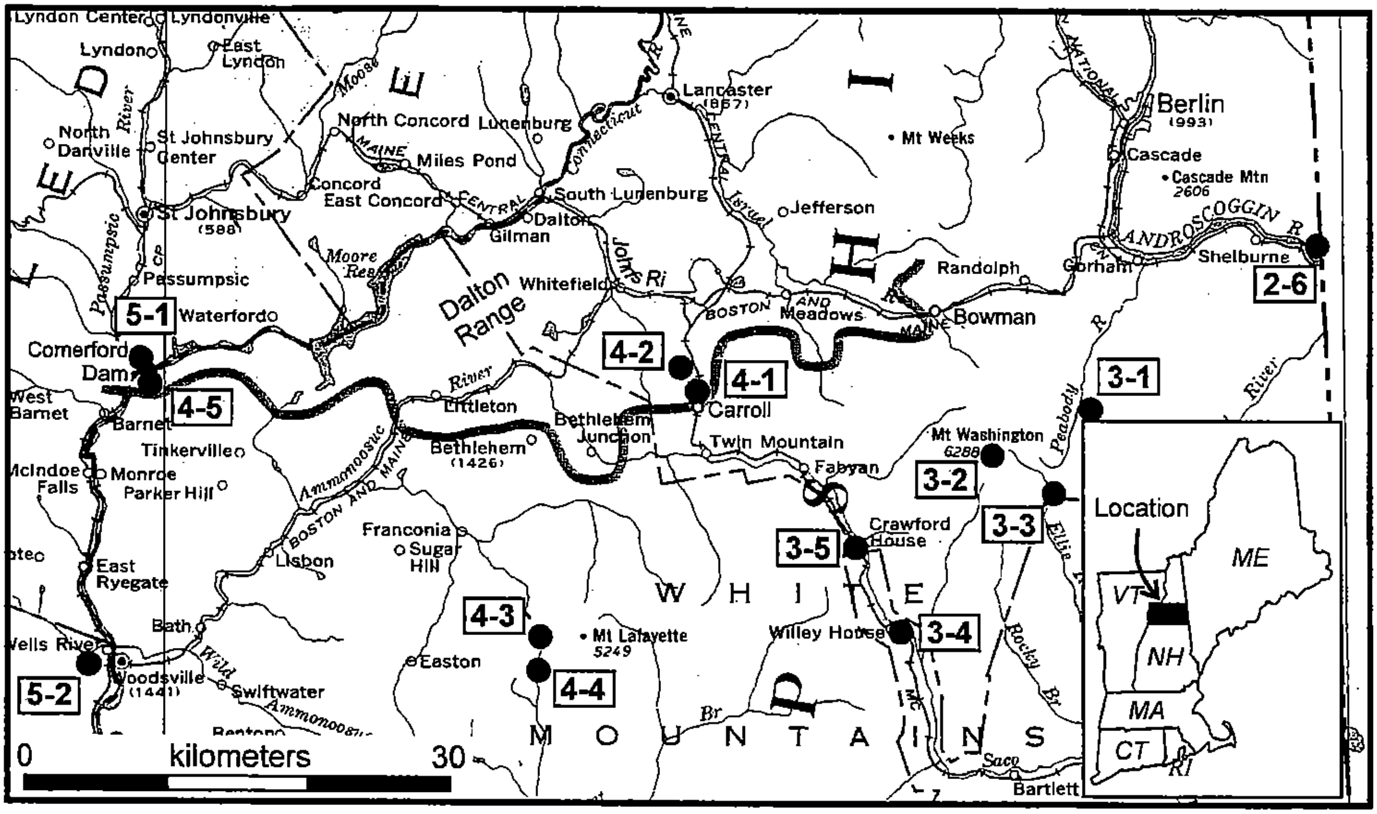

Figure 6. White Mt. area showing many of the stops for Days 2 through 5.

R.P. Goldthwait (1938, 1939, 1940, 1970a,b) carried on his father's interest in the glacial history of the Presidential Range. In his 1939 and 1940 publications, he not only noted the observations of his father in support of cirque glacier activity only preceding the last overriding by continental ice, but he also observed roche moutonnees on cirque floors along with striae and grooves on cirque headwalls, which he believed could only have been formed by continental ice (Davis, 1999). Also, in his latter paper, he presented morphometric data on cirques and altitudinal estimates of firn lines for the former cirque glaciers in the Presidential Range (Fig. 7). From these data, he calculated that, depending on the amount of winter precipitation, a 5 to 10°C mean summer temperature lowering would be necessary to support cirque glaciers in the Presidential Range today.

During the late 1950's, W.F. Thompson (1960, 1961) analyzed aerial photographs of the Presidential Range and refuted the Goldthwaits' view by arguing that the steep headwalls and sharp arêtes were indicative of active cirque glaciers following continental ice sheet deglaciation. Although Thompson (1960, 1961) did not present field data to support his view, he believed that moraines of cirque glaciers had been obliterated by post-glacial mass wasting processes (Davis, 1999). Work in Tuckerman Ravine by D.J. Thompson (1999) suggests that a deposit consisting of large blocks believed to be a moraine by Antevs (1932) is a relict, tongue-shaped, rock glacier unrelated to cirque glacier activity (Fig. 7).

Bradley (1981) challenged the Goldthwaits' view of the timing for cirque glaciation in the Presidential Range by noting that large boulders and diamicts at the mouths of north-facing cirques were composed of lithologies derived from bedrock in the cirques. However, Gerath and Fowler (1982), Fowler (1984), Gerath et al. (1985), and Waitt and Davis (1988) examined the diamicts at the cirque mouths and concluded that the sediments are not till, but rather debris flow deposits.

Opponents of post-ice sheet local ice support their views that continental ice was last with the following evidence: (1) roches moutonnees indicating upvalley ice flow on cirque floors, (2) striae trending obliquely across cirque headwalls, (3) ice sheet erratics in cirque–floor drift, and (4) apparent absence of moraines in or downvalley from cirques. Proponents counter that cirque glaciers did not completely remove ice sheet drift and that moraines in cirques are obscure because they are small, subdued by post-glacial mass wasting, or thickly forested. An important theoretical argument against post-ice sheet local ice in the Presidential Range is that, based on oxygen isotope and other proxies of late and immediate postglacial climate, such as pollen records, equilibrium-line altitudes rose rapidly to elevations well above cirque floors. Thus, cirque glaciers could not likely have been developed or sustained in the cirques following recession of continental ice.

Opportunities for developing a radiocarbon chronology for the deglaciation of cirques are limited because of the small num-

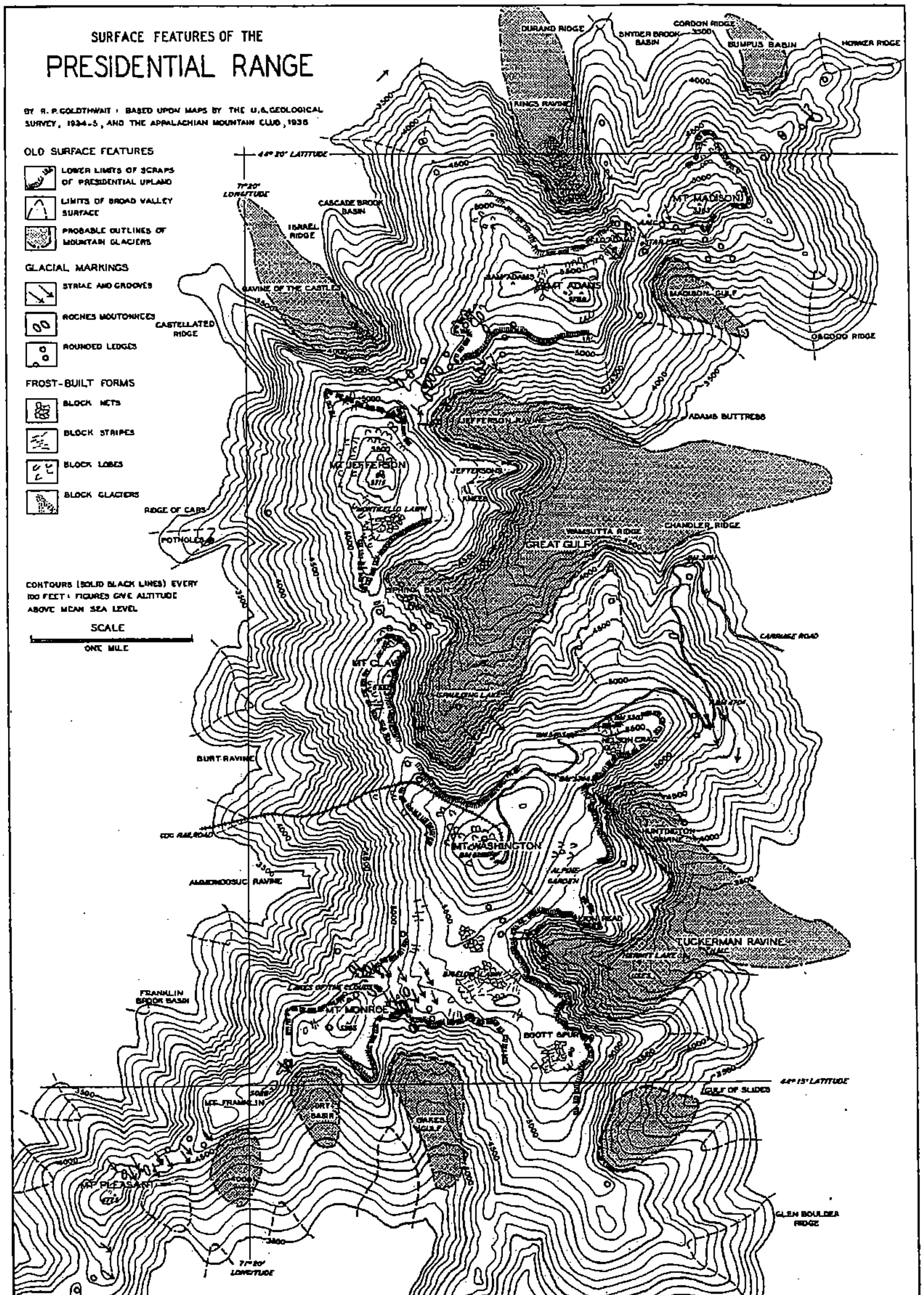

Figure 7. Surface features of the Presidential Range, including outlines of former cirque glaciers (from Goldthwait, 1940).

ber of tarns distributed across the range. Spaulding Lake in the Great Gulf and Hermit Lake in Tuckerman Ravine, although shallow, may provide useful continuous postglacial records of sediment accumulation and should be cored (Fig. 7).

Ponds near the cirques in the Presidential Range have provided minimum ^{14}C ages for continental ice retreat (Davis et al., 1980; Spear, 1989; Spear et al., 1994). Organic material from sediment at the base of a core retrieved from Lost Pond at an elevation 650 m in Pinkham Notch on the east side of the Presidential Range provided an age of 12,870 ± 370 ^{14}C yrs. B.P. (QL-985) (Spear et al., 1994). Organic material from sediments near the base of cores taken from the lower of the two Lakes of the Clouds at an elevation of 1542 m in the alpine zone between Mts. Monroe and Washington yielded an age of 11,530 ± 420 ^{14}C yrs. B.P. (I-10684) (Spear, 1989). Pollen data from sediments below the ^{14}C-dated level in the lower Lakes of the Clouds site correlate with the tundra pollen zone from Deer Lake Bog at an elevation of 1300 m on Mt. Moosilauke, which provides an age of 13,000 ± 400 ^{14}C yrs. B.P. (QL-1133) (Davis et al., 1980).

Given the model that continental ice thinned, separated, and retreated northward from the mountains of northern New England during late Wisconsin deglaciation (Goldthwait and Mickelson, 1982; Hughes et al., 1985; Stone and Borns, 1986; Borns, 1987; Davis and Jacobson, 1985; Thompson and Fowler, 1989), this entire process appears to have been very rapid. If these ^{14}C ages are taken at face value, they require almost 900 m of continental ice thinning in less than a few hundred years, a circumstance requiring very rapid warming of the local climate, warming too rapid to support local cirque glaciers. These ideas on the surficial geology of the White Mts. have been discussed on previous society field trips in the area (Davis et al., 1988; Davis et al., 1993; Davis et al., 1996; Allen et al., 2001).

Current work designed to refine the deglaciation chronology for the Presidential Range uses cosmogenic radionuclides ^{10}Be and ^{26}Al produced in quartz from boulders and bedrock. These exposure dating techniques (Bierman, 1994; Gosse and Phillips, 2001) may not provide the temporal resolution of AMS ^{14}C dating, but the method directly dates moraines and allows samples to be collected from sites where ^{14}C-datable materials are not available. As a test of the thinning continental ice model for deglaciation of the Presidential Range, a suite of bedrock and boulder samples with quartz veins were collected on an altitudinal transect from the summit of Mt. Washington to the floor of Pinkham Notch near Lost Pond for cosmogenic nuclide dating. The abundance of ^{10}Be and ^{26}Al in frost-riven bedrock samples from near the summit area of Mt. Washington is much higher (1.5 to 8 times) than expected had the peaks been covered by active, erosive ice during the late Wisconsin maximum (Bierman and Davis, 2000). Two summit samples provide ^{10}Be ages of 124,000 and 22,000 years exposure since deglaciation or the last active ice erosion. In contrast, a sample from one of the large boulders on the tongue-shaped rock glacier on the floor of Tuckerman Ravine (D.J. Thompson, 1999) appears to be late-glacial (Bierman and Davis, 2000). The mountain-top samples are consistent with two different scenarios, both of which have significant implications for understanding the spatial and temporal patterns of glaciation and glacial erosion in northern New England: (1) late Wisconsin continental ice was thinner than previously supposed leaving the tops of Mt. Washington's summit exposed since marine isotope stage 6, or (2) the summit was covered by glacial ice during the late Wisconsin, but the ice was thin enough to be frozen to its bed (<100 meters thick). Thus, the cold-based ice was unable to erode much rock, allowing cosmogenic nuclides to be inherited from prior periods of exposure. In either case, late Wisconsin continental ice in New England was thinner than previously believed, consistent with low, basal-shear stresses and/or the drawdown of continental ice by active ice streams flowing through the notches ("notch" is a local name for glacial U-shaped valley).

STOP 3-1. CIRQUES OF THE NORTHERN PRESIDENTIAL RANGE, PINKHAM NOTCH, N.H.

Mt. Washington (elev. 1950 m; 6450 ft.) is the highest peak in the Presidential Range and America's highest peak east of the Mississippi River and north of the Carolinas (Fig. 7). Mt. Washington also boasts the world's worst weather, its famous observatory having recorded the world's highest wind gust of 231 mph in 1934, along with other extremes of cloudiness, temperature, and precipitation. Pinkham Notch is located at ~650 m along the eastern base of Mt. Washington. Weather cooperating, the base of the Auto Road provides a superb panoramic view of the large east-facing cirques of the Presidential Range (Fig. 7). Glacially-eroded, bedrock forms that appear to be roches moutonnees occur nearby, but what do they indicate about ice-flow direction at this location?

Pinkham, Crawford, and Franconia Notches, from east to west, are the largest examples of glacial troughs in the White Mts., and all three are accessible from highways. Other smaller, but no less spectacular, notches in the White Mts. include, from east to west, Grafton, Mahoosuc, Carter, Carrigain, Zealand, and Kinsman Notches, some of which are only accessible by hiking trails. Each of these notches was a low-elevation mountain pass prior to the start of the multiple episodes of Pleistocene continental glaciation. The three notches that we will visit on this trip are substantially deeper than other higher elevation notches in the White Mts. because glacial ice moved through them for a longer period of time and/or was more erosive during each glacial episode. Also, the notches were likely locations of ice streams within ice sheets when the White Mts. were completely inundated by ice.

Looking northwest of the White Mts. highland area, continental glacial ice had few topographic obstacles interfering with its movement southeastward until it got to the steep northwesterly slopes of the White Mts. Because continental ice sought the path of least resistance through the mountains, it converged first into the deeper pre-existing passes through the mountains and continued to flow through them for the entire duration of each glacial episode. As continental ice thickened during each glacial episode, it began to move through the higher–elevation passes and eventually covered all the peaks at least once.

We know that all of the peaks here were over-ridden at least once by continental ice because lodgment till has been found near the summit of Mt. Washington at an elevation of 1,900 m (6270 ft.). Most workers agree that the last pulse of the late-Wisconsin Laurentide Ice Sheet arrived here about 25,000 ^{14}C yrs. B.P., reached its peak about 18,000 ^{14}C yrs. B.P., and had fully retreated from lowlands in the area by about 12,000 ^{14}C yrs. B.P. Thompson et al. (1999) have provided a detailed review of ^{14}C ages for the region north and northwest of the notches. However, because many of these ^{14}C ages are on bulk disseminated organic material, Ridge et al. (2001) question their accuracy, and based on an AMS ^{14}C-controlled varve chronology propose that deglaciation in the White Mts. was hundreds of years younger than the oldest ages summarized by Thompson et al. (1999).

On face value the ^{14}C ages from bog and pond bottoms in the White Mts. might suggest that deglaciation and organic sediment accumulation occurred earlier at lower elevations (ex. Lost Pond) than at higher elevations (ex. Lakes of the Clouds). However, palynological data (Davis et al., 1980; Spear, 1989; Spear et al., 1994; Cwynar and Spear, 2001) suggest that postglacial revegetation and organic sediment accumulation in bogs and ponds occurred at nearly all elevations at roughly the same time. Hence, the pollen data suggest that regional climate warmed rapidly during the early post-glacial, which supports the argument that glacial equilibrium-line altitudes rose too quickly to allow cirque glaciers to exist during the post-glacial.

STOP 3-2. SUMMIT, MT. WASHINGTON, N.H., LUNCH STOP

We will drive the 19 km (7.6 mi) Auto Road to the summit of Mt. Washington, making one or two stops on the way up to afford views of the Great Gulf cirque complex on the eastern side of the Northern Presidential Range (Fig. 7). On the summit, we will tour the Mt. Washington Observatory, the longest continuously operating mountain summit observatory on Earth. We will also examine sites where bedrock and frost-rived blocks sampled for cosmogenic nuclide exposure dating suggest that the summit was not overridden by erosive glacial ice during the last glacial maximum (LGM).

STOP 3-3. GLEN ELLIS FALLS, PINKHAM NOTCH, N.H.

A plaque above the falls titled "Geology of Glen Ellis Falls" reads "The Ellis River which had flowed uninterrupted during pre-glacial times was forced by violence and the struggle of the landmasses to plunge over the headwall of a glacial cirque." Not true! No cirque occurs here. The Cutler River, which feeds the falls, only recently adopted its present course (Fig. 6). Glen Ellis Falls is here because a major silicified zone, likely a Mesozoic fault with only a few meters of offset, juxtaposes the calc-silicate Wildcat granite or Rangeley Fm. migmatite against the Smalls Falls Fm. (D. Eusden, oral communicatioin, 2002). The 3-m-wide zone of massive quartz with many cross-cutting quartz veins is quite resistant and forms the escarpment over which the falls plunge. The silicified zone can be followed for about a mile in total, up and downstream from the falls.

STOP 3-4. WILLEY HOUSE SITE, CRAWFORD NOTCH, N.H.

Both walls of Crawford Notch (Mt. Webster on the east side and the Willey Range on the west side) are heavily scarred by landslide tracks (Figs. 6, 8). The largest landslide scar on the west side of the notch occurs on Mt. Willey, the site of the infamous Willey slide that occurred at 3 a.m. on Monday, 28 August 1826. As the story goes, upon hearing the roar of the slide, all family members ran out of their house, only to be obliterated while the house survived. Apparently a large boulder sitting directly upslope split the slide into two paths around the house. Following a hot and dry summer, landslides occurred all over the White Mts. during this two–day storm in 1926 (Silliman et al., 1829; Ramsey, 1988).

STOP 3-5. CRAWFORD HOUSE SITE AND "THE GATEWAY," HEAD OF CRAWFORD NOTCH, N.H.

We will walk east along the railroad tracks to a spectacular view down Crawford Notch. The steep slope at the head of the notch is not a cirque headwall at this low altitude, just as the precipice at Glen Ellis Falls in Pinkham Notch is not a cirque headwall (Figs. 6, 8). This valley's nick point probably predates Quaternary glaciation, and its presence here may result form harder bedrock types, such as the hornfels of Silurian Rangeley gneiss in the contact aureole of an intrusion of the Jurassic Conway granite. The darker crest of the ridge on Mt. Webster on the east side of the valley is also composed of the Rangeley and marks the roof of the Conway granite intrusion. This contact is exposed at road level on US 302 about 0.9 mi. east from the head of the Notch.

The head of Crawford Notch also was a spillway for glacial Lake Crawford to the west, which was contemporaneous with glacial Lake Pigwacket in the Conway area south of Crawford Notch (Thompson, 1997). The voluminous glacial meltwater that flowed down Crawford Notch during late-glacial time was orders of magnitude greater than postglacial runoff, and was responsible for the smaller V-shaped incision superposed on the classic U-shaped trough.

DAY 4. GLACIAL GEOLOGY OF THE LOWLANDS WEST OF MT. WASHINGTON AND FRANCONIA NOTCH, N.H. (Thompson, Davis, Fowler, and Ridge)

BETHLEHEM MORAINE COMPLEX AND ICE RECESSION IN THE NORTHWESTERN WHITE MTS., N.H. AND VT.

As originally mapped by J.W. Goldthwait (1916a), the Bethlehem moraine consists of dozens of hummocks and ridges in the Ammonoosuc River basin from Littleton east to Bethlehem (Fig. 6). The moraine ridges trend east to northeast, range in height from 3 m to over 30 m, and are up to 1300 m long. Few

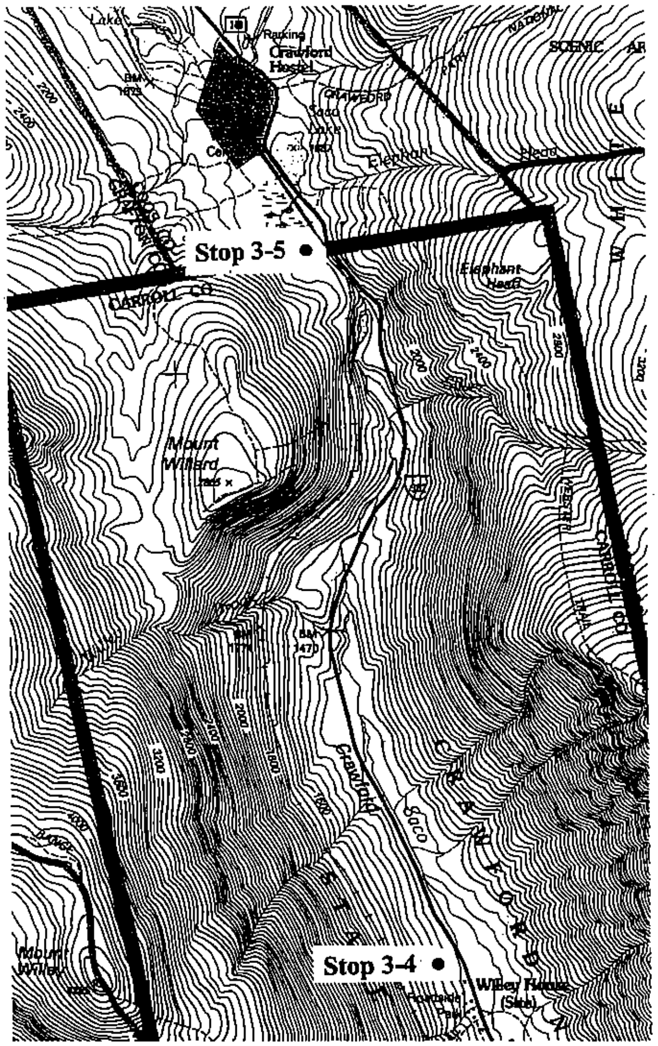

Figure 8. Part of the Crawford Notch, N.H., 1:24,000 quadrangle, showing locations for Stops 3-4 and 3-5 and contours indicating classic U-shape valley. Crawford Notch State Park boundary shown with shaded line.

bedrock exposures occur within the Bethlehem moraine complex, but striated outcrops immediately to the north and south indicate ice flow directions of 175-185° (Goldthwait, 1916a; W.B. Thompson, unpublished data). Surficial deposits in the moraine complex are generally thick (up to at least 50 m). The moraine ridges are composed of silty, sandy till and minor waterlaid sediments. Large boulders of granite and gneissic granite are abundant on the moraine surfaces. The elevations of the ridges suggest that they were deposited in glacial Lake Ammonoosuc, an interpretation supported by deformed glaciolacustrine sediments within the moraines.

The moraines of the Bethlehem area have been the focus of major controversies involving the modes of glaciation and ice retreat in the White Mts. (Thompson, 1999). Early workers accepted them as moraines deposited by active ice, but in the mid-1900s, ice-stagnation proponents opposed this concept. Thompson et al. (1999) re-examined the Bethlehem moraine complex and correlative deposits to the east. They concluded that the numerous subparallel till ridges are in fact end moraines deposited by oscillatory northward retreat of the ice margin during late Wisconsin deglaciation.

Thompson et al. (1996, 1999, 2002) compiled ^{14}C ages from ponds in the White Mts. and adjacent areas. Most ages are from the lower or basal portions of pond sediment cores and thus limit the time of deglaciation of each site. Pond of Safety, in the mountains just north of the Presidential Range, is favorably located to provide a minimum age for deposition of the Androscoggin moraine to the east and a maximum age for the Bethlehem moraine complex to the west. The basal age from Pond of Safety is 12,450 ± 60 ^{14}C yrs. B.P. (OS-7125), which is compatible with other limiting ages from northern N.H., as well as an age of 12,250 ± 55 ^{14}C yrs. B.P. (OS-7119) from Surplus Pond in western Maine. The cores from both ponds show sedimentological evidence of a climatic reversal that probably occurred in the Younger Dryas (Thompson et al., 1996).

Additional ^{14}C ages were recently obtained from cores taken in other ponds in the northern White Mts. (Boisvert et al., 2002). Together with previous results, the new data suggest that the Littleton-Bethlehem Readvance (and deposition of the Bethlehem moraine complex) occurred ~ 12,000 ^{14}C yrs. B.P. (~ 14,000 cal. yrs. B.P.). A short distance west of Littleton, Ridge et al. (1996, 1999, this volume) found stratigraphic evidence supporting the hypothesis of previous authors that a glacial readvance occurred in the vicinity of Comerford dam on the Connecticut River (Antevs, 1922; Crosby, 1934a; Lougee, 1935). The latter authors equated this readvance with the one that deposited the Bethlehem moraine complex. Lougee (1935) referred to this event as the "readvance at Littleton," and Thompson et al. (1999) named it the "Littleton-Bethlehem Readvance" to emphasize the correlation with the deposits historically known as the Bethlehem moraine.

Based on recent work on the glacial Lake Hitchcock varve sequence in the Connecticut Valley, Ridge et al. (1999) determined an age of 11,900-11,800 ^{14}C yrs. B.P. for the Littleton-Bethlehem Readvance (see also STOPS 5-1 to 5-3). Moreover, Larsen (2001) obtained an age of 11,900 ± 50 ^{14}C yrs. B.P. for the Middlesex Readvance near Montpelier, Vt., which he correlated with the readvance in the Comerford Dam area (and presumably with the Littleton-Bethlehem Readvance in general). These findings suggest that the Littleton-Bethlehem Readvance and its correlatives to the east and west were a major glacial event across this part of northern New England. Thompson (1998) proposed that the Littleton-Bethlehem Readvance occurred during the Older Dryas. This cold interval began ~ 12,200-12,000 ^{14}C yrs. B.P. and lasted only about 200 years (Donner, 1995; Wohlfarth, 1996). The GISP2 ice core from Greenland likewise indicates a brief cold period ~ 12,000 ^{14}C yrs. B.P. that is equated with the Older Dryas (Stuiver et al., 1995). Ridge's work in the Connecticut Valley varve sequence (STOP 5-3) further supports the Older Dryas age for the Littleton-Bethlehem readvance. If deposition of the Bethlehem moraine complex were limited to such a brief interval, the ice sheet must have responded quickly in northern N.H.

STOP 4-1. TWIN MT. SAND & GRAVEL PIT, CARROLL, N.H. (Thompson)

Glacial lake deposits are among the most useful features for determining the mode and sequence of deglaciation in the northern White Mts. Ice-dammed lakes existed in north-and west-draining valleys in this area, and most of the Bethlehem moraine complex was deposited in a lake that occupied the Ammonoosuc Valley. Goldthwait (1916a) named this water body "Lake Ammonoosuc," which resulted from damming of the valley when the late Wisconsin ice margin stood in the Littleton-Bethlehem area. As the margin receded, successively lower spillways for the lake were uncovered and the lake level fell. The earliest stage of the lake (Crawford Stage) drained eastward through Crawford Notch, where the spillway has an elevation of about 573 m (1880 ft).

Following the Crawford Stage, glacial Lake Ammonoosuc drained southwestward through five progressively lower spillways into Gale River valley in the Franconia area (Fig. 9). The spillway for the Gale River 2 stage is a prominent channel that will be seen during this trip along Route 3 southwest of Twin Mt. Later spillways north of Bethlehem village drained the Bethlehem and Wing Road Stages of Lake Ammonoosuc into Indian Brook and finally into the lower Ammonoosuc River (Lougee, n.d.; Thompson et al., 1999).

The large pit at this stop is located in the ice-contact "Carroll delta" (Lougee, 1940), which built southward into the Gale River 2 stage of glacial Lake Ammonoosuc. This deposit formed when the ice margin was pinned against Beech Hill to the west and Cherry Mt. to the east. Figures 9 and 10 show the relationship of the Carroll delta to meltwater channels and ice-margin positions. A short esker segment along the railroad track north of the pit marks a feeder channel for the delta. Additional meltwater and sediment came from channels on the hillside to the northeast. The lake level dropped when recession of the ice tongue blocking the lower Ammonoosuc Valley opened up the next lower spillway (G3 on Fig. 9). The resulting incision of the Carroll delta formed the prominent channel seen along the railroad track. The water

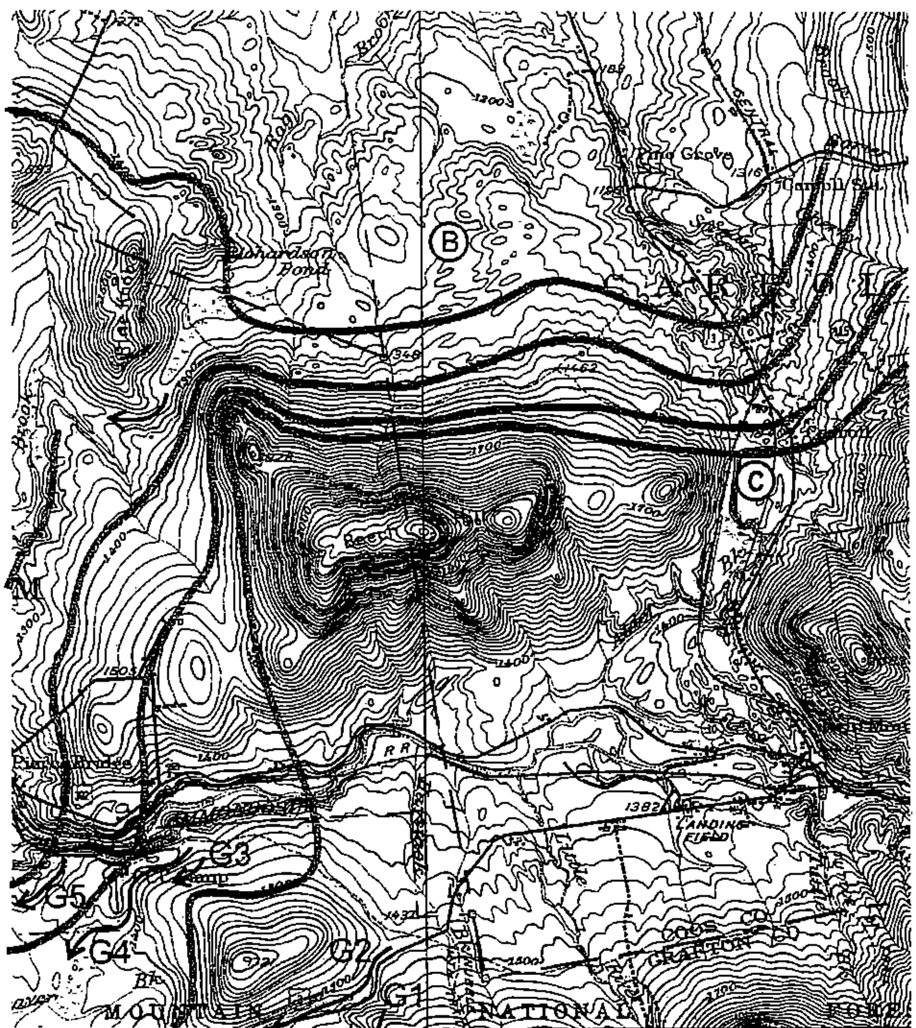

Figure 9. Part of the Whitefield, N.H., 1:62.500 quadrangle, showing inferred ice-margin positions (gray lines) and meltwater spillway channels (arrows). G1-G5 are spillways for successive Gale River stages of glacial Lake Ammonoosuc. B: Beech Hill moraines (STOP 4-2). C: Carroll delta (STOP 4-1). (From Thompson et al., 1999).

that cut the channel may have come initially from the more northerly of ice-margin positions shown in Figure 10, and later from glacial Lake Carroll to the north.

Remnants of coarse gravel forming the delta topset beds can be seen along the upper east wall of the pit. In 2000, the foreset beds were best exposed in a newly deepened area of the pit just east of the processing plant. The southern part of this opening showed many thrust faults in the delta foresets, while the west side (closer to former ice margin) revealed bouldery, ice-contact gravel and till. The faulting indicates ice shove against the back-side of the delta, which is further evidence of late-glacial ice activity in the region.

STOP 4-2. BEECH HILL MORAINES, CARROLL, N.H.
(Thompson)

The Beech Hill moraines (Fig. 9), at elevations of 396-427 m (1300-1400 ft), are till ridges that are 4-14 m high, up to 700 m long, and trend ENE-WSW. A small borrow pit in one of the ridges shows sandy, light olive-gray till. Large (1-3 m) boulders of coarse white granite are extremely abundant on the surfaces of the moraines. The boulders were derived from the local bedrock, which is "moderately to weakly foliated biotite granite" of the Oliverian plutonic suite (Lyons et al., 1997). Many of the granite boulders are weathered or disintegrated. A great variety of rock types in the pit include igneous and metamorphic lithologies.

Meltwater channels occur on the distal sides of the Beech Hill moraines. The largest channel lies at the southern margin of this moraine cluster and hosts a wetland and small pond seen on the newest Bethlehem quadrangle map. The channels developed sequentially in the swales between moraine ridges and thus are parallel to the moraines. Meltwater flowed southwest along the ice margin as it retreated from Beech Hill and drained through the 387-m gap between the northwest corner of Beech Hill and Pine Knob (Fig. 9).

Thompson et al. (1999) correlated the Beech Hill moraines with the Bethlehem moraine complex to the west and moraines in the upper Israel River valley to the northeast. They attributed all of these moraines to the Littleton-Bethlehem readvance. Many other moraines occur in the area between this stop and Whitefield village.

STOP 4-3A. VIEW OF OLD MAN OF THE MT., FRANCONIA NOTCH, N.H., LUNCH STOP
(Davis and Fowler)

Exit I-93 at the "Old Man Viewing" ramp sign in Franconia Notch State Park parking lot and walk ~ 500 meters to the view-point on the north shore of Profile Lake (Fig. 6).

The profile is made up of six fortuitously-broken, plate-shaped blocks of bedrock positioned above each other in such a way as to create the distinctly "human profile" best seen from the shore of Profile Lake, 550 meters below (Fig. 11). The Profile is about 21 meters high and is estimated to weigh between 10,000 and 13,000 tons. The six blocks or plates comprising the Profile represent, from top to bottom: (1) forehead, (2) eyebrow, (3) nose, (4) base of nose and upper lip, (5) lower lip and chin, portion separated from the cliff, and (6) low lip and chin, portion behind No. 5 and not separated from the cliff. These blocks are vertically separated from one another along sub-horizontal joints (N25°W, 23°NE) that are part of a single set of joints. The Profile itself has been created by selective breakage at the edge of these blocks along nearly vertical joints (S65°W, 73°SE), the predominant set forming Cannon Cliff; N20°E, 75°SE; N35°E, 80°SE; N60°W, 60°SW; N25°E, vertical; N20°W, 85°NE)(Fowler, 1982, 1997).

The information presented here is from a study conducted in 1976 as a part of the environmental impact statement analyzing alternatives for Interstate-93 through Franconia Notch (Fowler, 1982). The study concluded that limited blasting could be undertaken if closely monitored, and it went on to draw the following conclusions about the stability of the Profile. (1) The principal mechanism responsible for the Profile's stability is the compressive cantilevering created by the dead weight load of most of the blocks comprising the Profile acting at the location of their combined center of gravity behind the cliff-face, (2) The blocks in the rock mass that are the most stable are those from the nose upward to the forehead because their large intact masses and the location of the individual and combined centers of gravity well behind the cliff face act to enhance the cantilevering mechanism, (3) The blocks in the rock mass that are the least stable, and which pose the most important stability problems, are those from the upper-lip downward to the chin because their comparatively small intact masses and the location of their individual centers of gravity very near to or (in the case of the chin) in front of the cliff-face are not part of the overall cantilevering mechanism and they obtain no support from the blocks that comprise the rest of the profile. (4) Because of the positions and weights of the blocks in the profile, the active mechanical reinforcement of the various portions of the rock mass to increase their security appears possible, but additional structural and mechanical analyses are needed to design a system that can be installed without endangering the profile in the process.

STOP 4-3B. PROFILE LAKE, LANDSLIDES AND ROCKFALLS, FRANCONIA NOTCH, N.H.
(Davis and Rogers)

Walk about 100 m south of the Old Man viewpoint to a point overlooking Profile Lake and a view of the spectacular slopes of Franconia Notch to the south (Fig. 11). The broad expanse of Cannon Cliff dominates the western slope and the bulky shoulder of Mt. Lafayette looms over the eastern slope of the notch. Mt. Lafayette is composed of porphyritic quartz syenite. Cannon Cliff is composed of Conway granite, which exhibits textbook examples of exfoliation slabs, especially striking at the north end. A deep recess at the south end of the cliff owes its origin to a thick diabase dike, which trends roughly NNE through the higher notch separating Eagle Cliff and Mt. Lafayette on the eastern

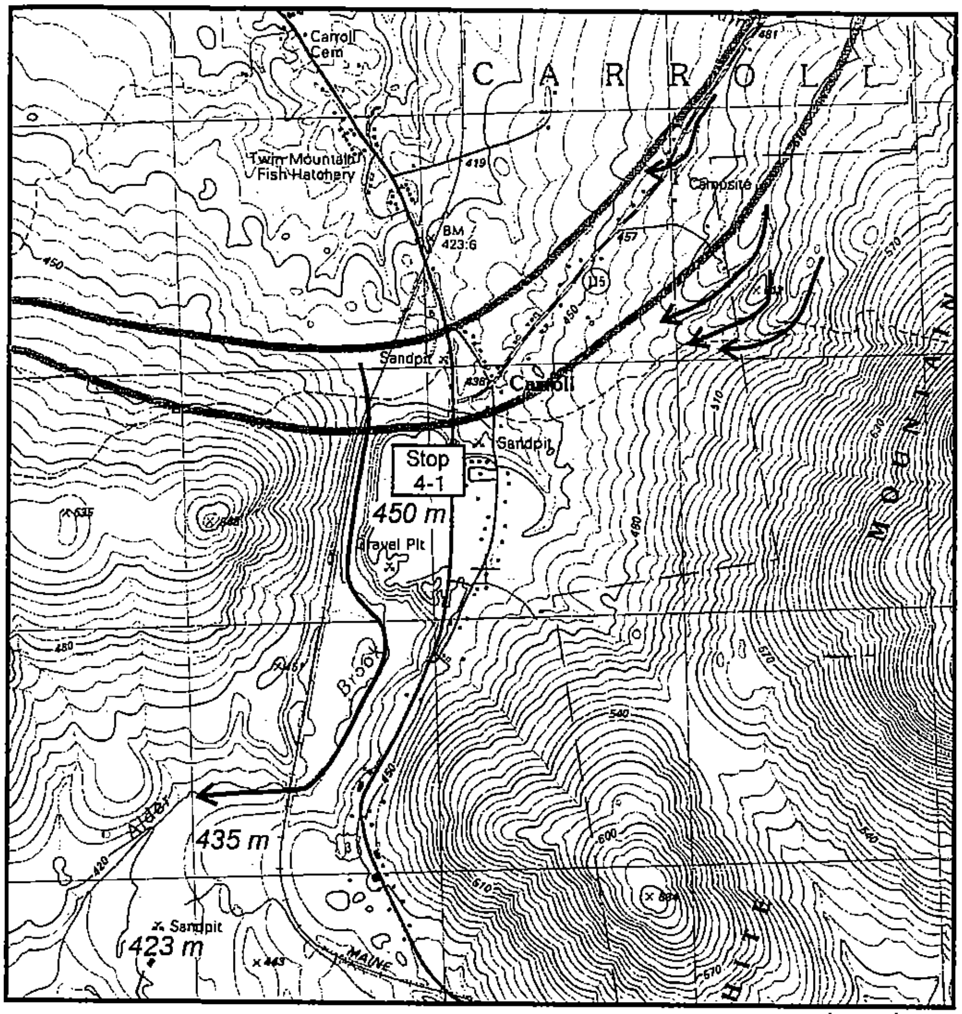

Figure 10. Part of the Bethlehem, N.H., 1:25,000 metric quadrangle, showing the location of the ice-contact Carroll delta at elevation of 450 m (STOP 4-1); the 435- and 423-m surfaces graded to lower stages of glacial Lake Ammonoosuc; inferred ice-margin positions (gray lines); and meltwater channels (arrows).

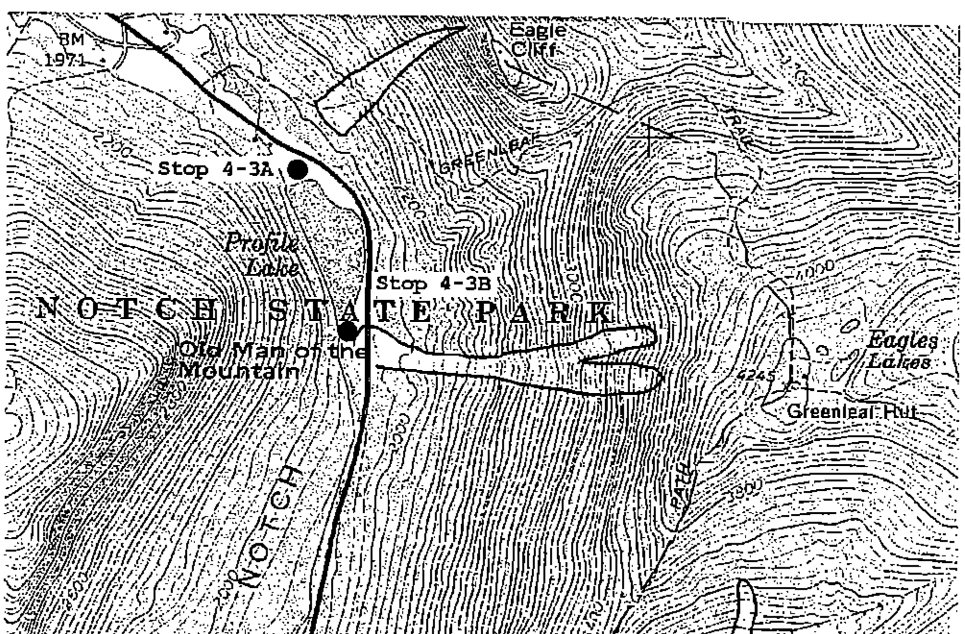

Figure 11. Part of Franconia Notch, N.H., 1:24,000 quadrangle, showing locations for STOPS 4-3a and 4-3b. Solid line is I-93 parkway. Note landslide scars on the east side of valley adjacent to Profile Lake. STOP 4-4 is located about 5 km south of Profile Lake on the west side of parkway.

slope of the Notch. Cannon Cliff is one of the East's premier destinations for rock and ice climbing. The south end of Profile Lake is dammed by landslide debris delivered from the west-facing slope of Mt. Lafayette in the area of the prominent landslide scar (Fig. 11). This form of mass wasting has been a very important mechanism in the modification of the landscape in Franconia Notch and other notches in the White Mts. since deglaciation.

As the regional climate rapidly warmed when continental ice was receding to the north, the newly-exposed bedrock surfaces of the mountains and valleys were attacked by the harshly variable climate that existed close to the retreating glacier. This climate was characterized by large and frequent differences in diurnal temperatures, which permitted intense freezing and thawing of water in cracks and crevices in the bedrock surfaces. This alternating freezing and thawing shattered the bedrock and created large volumes of blocky talus on the mountain peaks and slopes. As this talus continued to accumulate, it either slid or fell to the bases of the cliffs and valley walls, forming the Notch's ubiquitous talus slopes. This process lessened in intensity as the local climate continued to warm and stabilize, so the amount of shattering and rockfall has decreased but not stopped altogether.

The large talus slope at the base of the Cannon Cliff (Fig. 11) is an excellent example of rockfall debris accumulation and although rockfall is not as intense today as it was 10,000 to 12,000 years ago, it is still continuing and poses a considerable safety threat to rock climbers on the cliff and even those scrambling up to the base of the cliff. The now-sharply-pointed peaks of Franconia ridge (from north to south, Mts. Lafayette, Lincoln, Little Haystack, Liberty, and Flume) are the result of their glacially rounded summits being shattered and sharpened by the freeze-thaw process. Many of the alternative construction designs for Rt. I-93 were abandoned because of the expense and difficulty associated with construction in such unconsolidated materials (e.g. cut-and-cover tunnel). Finally, in the early 1980s, I-93 was completed as a parkway, and is the only non-four-lane section of the interstate highway system in the U.S.A.

Highway crews frequently clean up landslide debris from the bottom of the many slide tracks that scar all three major Notches visited on this trip. However, Franconia Notch has been particularly problematic. Numerous historic landslides have occurred on Eagle cliff and the western flank of Mt. Lafayette (Flaccus, 1958), with major events occurring in 1826, 1850, 1883, 1915, 1947, 1948, 1959, and 1974. The landslides of 1948 and 1959 covered old US 3 with so much debris at this location that many days were required to re-open the highway. Many of these landslides were caused by extreme rainfall events.

During the winters of 2000 and 2001, a research project was initiated to recover sediment cores from Profile Lake in order to reconstruct a pre-historic record of large landslides in Franconia Notch (Rogers et al, 2001; Rogers, 2003). Samples dated from near the base of the longest cores from the middle of the lake suggest that sediment accumulation began over 10,000 ^{14}C yrs. ago. From near the tops of sediment cores at the south end of Profile Lake, layers of coarse debris that fine upwards (based on detailed grain size analysis by Sedigraph) (Rogers, 2003), thicken toward the east shore of the lake. Magnetic susceptibility (MS) and loss-on-ignition (LOI) measurements allow correlation of coarse layers in near-shore, sediment cores with clastic lenses of finer grain-size in cores from the middle of the lake. Measurement of ^{210}Pb and ^{137}Cs profiles in the long sediment core tops from the middle of the lake suggest that the uppermost clastic lens reflects the 1959 landslide that covered the highway and extended into the south end of the lake. This analog for a major landslide triggered by an intense rainstorm (ex. Sharpe, 1938) is used to reconstruct a pre-historic record of landslides, which may indicate extreme rainfall events since deglaciation over 10,000 years ago. The sediment record from Profile Lake supports the work of Bierman et al. (1997), whose study of alluvial fans and pond sediments suggested that hillslope erosion in northern Vermont was caused by severe storms during the early and late Holocene. However, the Profile Lake landslide record does not support millennial-scale storminess variability as recorded by the compilation of sediment core data from 13 lakes in Vermont and eastern New York (Noren et al., 2002).

Cannon Mt. has one of the highest relief cliffs in northeastern United States (~300 m). At about 10 a.m. on June, 19, 1997, a large rockfall originated from near the top of the cliff's north end (elevation ~975m). The release involved a large section of steeply inclined exfoliation slabs of Conway granite, which became airborne over the lower half of the cliff. The rockfall was apparently not triggered by seismic, sonic, or meteorological events, as observers did not note an earthquake, sonic boom, or quarry blast, and the previous four days were oddly precipitation-free. Both the cliff scar and the impact swath on the talus slope below the cliff remain clearly visible from the valley floor.

Cannon Cliff's talus slope, with a relief of about 250 m, is composed of angular blocks generally <5 m in length. Many parts of the talus slope are vegetated with trees and shrubs, but in open areas most talus blocks are covered with gray to black lichens, in contrast to the green lichens on the cliff face. Because of the altitude of the cliff, frost action is severe and rockfalls are common, especially during the spring.

The June 1997 rockfall not only flattened all vegetation, including shrub birch and spruce, but mobilized most blocks for a width of about 100 m on the upper part of the talus slope, leaving immense exposures of fines (less than pebble size), which have begun to erode since June 1997. The talus slope absorbed most of the energy of impact, but many rockfall blocks traveled up to 425 m, cutting a 10–20 m wide, 5–10 m deep trough into the lower part of the slope. Beech and maple trees up to 45 cm DBH were splintered or sheared. One large rockfall block (5.1 m x 3.5 m x 3.3 m; ~3.5 tons) reached the valley floor, long axis oriented upslope, adjacent to a paved bicycle path ~380 m south of Profile Lake. This block originated from the cliff face, as part of the top surface remains mostly unscathed and covered with green lichens. The arrangement of blocks comprising the Old Man of the Mountains is susceptible to the same mass wasting processes demonstrated by the 1997 event, mechanisms that no doubt will be responsible for the Old Man's ultimate collapse, given its delicate state of stability.

STOP 4-4. THE BASIN POTHOLES, FRANCONIA NOTCH, N.H. (Davis)

Although not the largest and deepest in the White Mts., the potholes at the Basin are the most accessible. Conventional wisdom suggests that potholes are created by the abrasive action of sand and gravel in swirling eddies of meltwater streams. However, cases may be made for an entirely subglacial origin of potholes, as they have been described at the top of a bedrock ridge in Ontario (Gilbert, 2000) that requires overlying ice to provide water flow in such a topographic setting (i.e., via glacial moulins). The same case may be made to explain potholes at an altitude of over 1200 m along the Caps Ridge trail on the western flank of Mt. Jefferson in the north Presidential Range.

STOP 4-5. MILL BROOK SECTION, MONROE, N.H. (Ridge)

Three tills separated by sand, gravel, and bedded diamictons occur along Mill Brook at the south side of the Comerford dam (Fig. 6; Ridge et al., 1996; Thompson et al., 1999). The lower two tills were first described by Crosby (1934b) as lower and upper tills on the west side of the brook. The present exposure, farther south and on the brook's east bank, has an additional till at its top overlain by varved silt and clay. Crosby (1934a) interpreted his upper till (our middle till) to represent a late Wisconsin readvance correlative with the Littleton-Bethlehem moraines farther east (cf. Thompson et al., 1999). Our uppermost till appears to be a better candidate for this readvance. The entire section may be late Wisconsin. Alternatively, the lowermost till may be pre-late Wisconsin, but this is not supported by evidence of weathering, as at other places in New England (Koteff and Pessl, 1985; Newman et al., 1990).

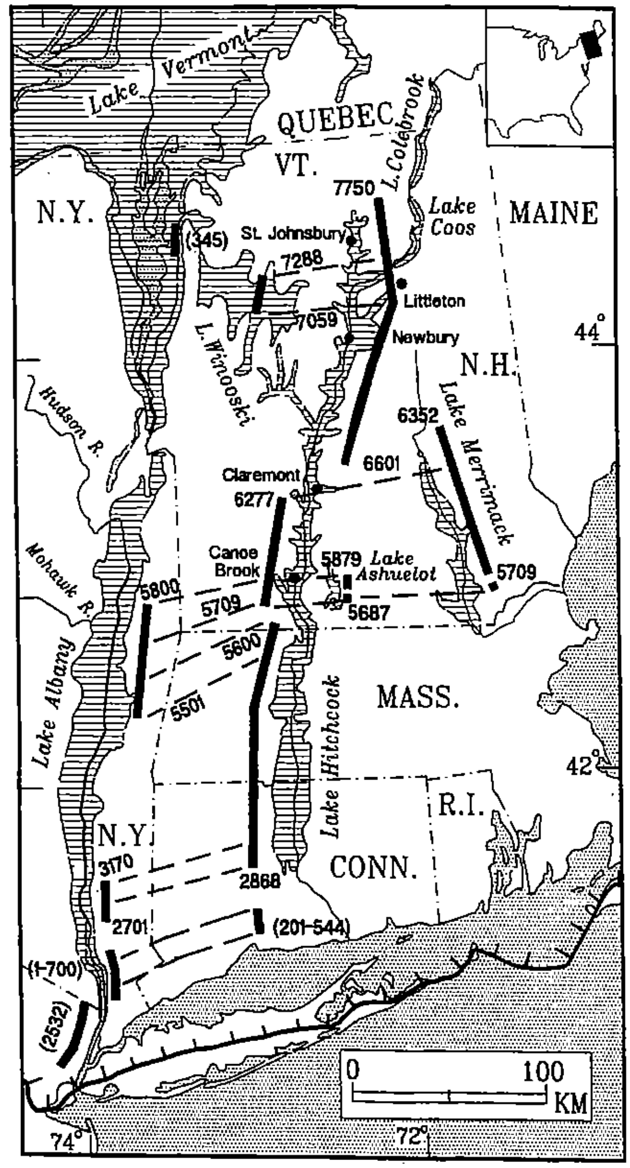

Figure 12. Important varve chronologies in the northeastern United States, including, from south to north, varve sequences from glacial Lake Hackensack in northern N.J. (2532 yr; Reeds, 1926; Antevs, 1928), correlated sequences from lake Quinnipiac near New Haven, Conn., and Lake Albany at Haverstraw, N.Y. (Antevs, 1928), correlated sequences (Antevs, 1922, 1928) of the lower (NE varves 2701-6352) and upper Connecticut varves (NE varves 6601-7750), and the Essex Junction, Vt., sequence (345 yr; Antevs, 1928).

DAY 5. CONNECTICUT RIVER LOWLAND, N.H.-VT. AND GREEN MTNS, VT.; THE NEW ENGLAND VARVE CHRONOLOGY (Ridge)

In the 1920's, Gerard De Geer and three of his students from the University of Stockholm, Ragnor Lidén, Ebba Hult de Geer, and Ernst Antevs, began work on varves in the northeastern United States to formulate a varve chronology similar to the Swedish varve chronology. However, Ernst Antevs (1922, 1928) almost single handedly formulated the two main sequences of the New England (NE) varve chronology (Fig. 12), in addition to sequences in the Hudson Valley, and N.J. with Chester Reeds (1926). Antevs began his work in southern Conn., with a varve sequence that arbitrarily starts at varve 2701, to accommodate older varves, and extends to varve 6352 (3652 varves). This varve sequence, the lower Connecticut varves, is mostly based on measurements in the Connecticut Valley from Hartford, CT., north to Claremont, N.H., and includes sequences from the Hudson Valley of N.Y., and Ashuelot and Merrimack Valleys of N.H. These additional sequences were critical to filling gaps and extending the chronology.

Antevs (1922, 1928) also compiled a younger sequence north of Claremont (Fig. 12), based on measurements from the upper Connecticut and Passumpsic Valleys of northern Vt. and

N.H., and the Winooski Valley of western Vt. This sequence is known as the upper Connecticut varves. It starts with varve 6601 and extends to varve 7750 (1150 varves) and does not have a clear overlap with the lower Connecticut varves. Antevs separated the sequences by what has become known as the Claremont Gap, which he interpreted to be the result of delayed ice recession that allowed deposition of the lower Connecticut varves while areas to the north remained ice covered. A crude similarity between the ends of the varve sequences, supported by similar paleomagnetic records and consistent ^{14}C ages, suggested an overlap of the varve sequences (Ridge et al., 1996, 1999), thus eliminating the Claremont Gap and implying continuous rapid deglaciation of the Claremont area.

An important step in using the varves as a chronologic tool has been the calibration of the NE varve chronology with ^{14}C ages so that both ^{14}C and calibrated (U-Th) time scales can be applied to the varves. This work began when ^{14}C ages were obtained from varve 6150 at Canoe Brook, Vt. (Ridge and Larsen, 1990). The calibration of the lower Connecticut varves appeared to be secure when consistent ^{14}C ages were also obtained from the upper Connecticut varves (Ridge and Toll, 1999; Ridge et al., 1999). However, the original ^{14}C ages from Canoe Brook were from the time of a plateau in the ^{14}C time scale (200 ^{14}C yr = 900 cal yr; 12.6-12.4 ^{14}C ka B.P.) that did not allow for a precise calibration of the lower Connecticut varves. A new ^{14}C age at Canoe Brook from varve 5858, (12,660±50, GX-25735), and from outside the ^{14}C plateau, has provided an improved calibration of the lower Connecticut varves. This new calibration makes the lower Connecticut varves older and based on ^{14}C ages they no longer appear to overlap the upper Connecticut varves (Ridge, 2003). Recent detailed mapping (Ridge, 1999, 2001) has also revealed three sets of end moraines associated with readvances in the Claremont area, indicating slowed ice recession and supporting a gap between the varve sequences (Ridge, 2003).

Along with the calibration of the NE varve chronology has been the refinement of paleomagnetic records from the varve sequence, especially a record of declination. Declination records were originally constructed by McNish and Johnson (1938), Johnson et al. (1948), and Verosub (1979a, 1979b). Recent work has focused on completing the declination record of the lower Connecticut varves and refining the upper Connecticut record with new laboratory techniques (Ridge et al., 1996, 1999, 2001). Correlation of the declination records from varves in New England and glacial lacustrine deposits in central N.Y. (Brennan et al., 1984; Ridge et al., 1990; Pair et al., 1994) has provided a means of testing ^{14}C ages and inferring correlations across the northeastern U.S. (Ridge et al., 1990, 1999, 2001; Ridge, 2003).

STOP 5-1. COMERFORD DAM SECTION, BARNET, VT.

At the lower end of the retaining wall embankment along the spillway for the Comerford dam is a gully exposure of varves resting on till (Fig. 6). Bedrock in the spillway channel is striated at 182°. The varve section is very close to a varve section on the spillway embankment studied by Lougee (1935) during the final phases of construction of the Comerford dam. He described a compact and deformed upper part of the varve section that was overlain by a "material resembling till." Lougee interpreted this till to represent the Littleton-Bethlehem Readvance. Lougee measured the varve exposure and matched the varve record to the New England varve chronology. The varve section at the present exposure has been measured (Ridge et al., 1996, 1999) and matches Lougee's measurements and the NE varve chronology (varves 7036-7154). However, the top of the varve section, with Lougee's overlying till unit, was truncated by completion of the retaining embankment. The basal varves resting on till allow a determination of the age of initial ice recession at this position in the Connecticut Valley. Further, if Lougee's (1935) observation of till at the top of the section is correct, the varve correlation, along with other nearby sections measured by Antevs (1922), precisely brackets the age of the Littleton-Bethlehem readvance.

STOP 5-2. VARVE SECTION, WELLS RIVER VALLEY, VT.

This exposure along the north side of the Wells River valley is a sand and gravel pit that exposes a thick section of varves from an embayment of Lake Hitchcock (Fig. 6). The sand and gravel is an esker that was buried beneath the varves. Several outcrops in the pit show the base of the varve section where it rests on sand and gravel, as well as the upper part of the varve sequence that grades upward into lacustrine sand. Many horizons in the pit show spectacular soft sediment contortions probably related to the instability of clayey sediment that was deposited on the flanks of the esker. The varves in the lower part of the section also contain concretions. Attempts to match varves in the upper stratigraphy with the NE varve chronology have failed and lower parts of the stratigraphy are presently being processed. Antevs (1922) also had trouble matching varve sections in this tributary valley. Local post-glacial depositional processes, related to meteoric or non-glacial runoff, may have masked annual weather patterns in the varves or may have recorded weather patterns differently than the glacially derived varves along the main axis of the Connecticut Valley to the east.

STOP 5-3. VARVE SECTION AND LUNCH STOP, NEWBURY, VT.

This section is along a small, unnamed stream that joins the Connecticut River at the outside of a meander just east of Rt. 5, about 15 km south of STOP 5-2. Antevs (1922) measured and matched 327 varves of the section to the NE varve chronology. He counted an additional 1184 varves above the measured section for a total of 1511 varves (varves 6990-8500 of Antevs, 1922, 1928). Most of the upper 1184 varves were too thin (some <1 mm) to measure realistically in the field with a ruler. In the 1940's, the Newbury section was again counted as part of one of the first studies in sedimentary paleomagnetism that recorded the secular variation of geomagnetic declination (Johnson et al., 1948). In the last 10 years, the entire section has been collected in duplicate PVC cores and measured with the aid of magnified video images and a computer measurement program (Ridge and Toll, 1999; Ridge et al., 1999). This recent analysis of the section has shown that it has 1735 varves spanning varves 6944-8679(+35/-20), after accounting for uncertainties in the interpretation of annual layers.

The lower Newbury section (varves 6944-7470) records deglaciation in the upper Connecticut Valley (Fig. 13). This part of the section has thick varves (mostly >1 cm) that are dominated by sediment from glacial runoff. The varves contain ostracods and the sinusoidal trails of nematodes. Scattered thick varves (more than an order of magnitude thicker than surrounding varves) are interpreted to be flood deposits formed by the release of water from tributary lakes impounded by the receding glacier. Most notable are the exceedingly thick couplets (up to 15 cm in varves 7200-7220) produced by the catastrophic release of water from lakes in the upper Ammonoosuc Valley about 50 km to the northeast (see STOP 4-5). These events do not appear in correlative records of the Winooski Valley. The implied source of the floods is based on the correlation of ice front positions in the Ammonoosuc and Connecticut Valleys at the Comerford dam, which is bracketed by the varve chronology (see STOP 5-1). The flood events mark the earliest recession of ice of the Littleton-Bethlehem Readvance. The period of readvance itself (varves 7115-7200), based on the Comerford dam correlation, is represented by relatively thin (< 1 cm) glacial varves that do not have erratic jumps in thickness representing floods. The period of ice advance appears to have been a time of reduced glacial meltwater production and the impounding, rather than catastrophic release, of tributary lakes by advancing ice.

At varve 7470, the varves abruptly become very thin (<5 mm), which represents recession of the glacial meltwater source and the start of non-glacial deposition in a paraglacial and periglacial environment. This non-glacial lake persisted in the Connecticut Valley for at least 1200 years after glacial runoff no longer contributed a significant volume of sediment to the lake. The non-glacial varves contain abundant fossil plants, ostracods of several species, and nematode trails. Two types of bivalves have been recovered including many *Sphaerid* clams and a single mussel specimen, *Pyganodon fragilis*, which is a cold-water species no longer found in New England (Smith and Ridge, 2001). Additional fossils include an as-yet-unidentified gastropod and sinusoidal, fish-swimming trails.

The plant fossils from the Newbury section, especially wood and woody twigs from terrestrial species, have yielded ^{14}C ages that provide a calibration of the upper Connecticut varves (Ridge et al., 1999; Ridge, 2003). Six ^{14}C ages from Newbury and an additional age from the Passumpsic River valley to the north, ranging 11.5-10.1 ka, are consistent with their positions in the varves and provide a calibration of the sequence to ^{14}C and calibrated (U-Th) time scales (Fig. 14). This calibration has allowed a comparison of the Newbury varve record with oxygen isotope records from Greenland ice cores (Ridge and Toll, 1999; Fig. 5-3a). The comparison indicates that the Littleton-Bethlehem Readvance corresponds to the Older Dryas. In addition, the Killarney Oscillation in North America (Intra-Allerød Cold Phase in Europe) appears to correspond to an interval of thicker varves that have a conspicuous 22-yr periodicity. Near the top of the non-glacial sequence the onset of the Younger Dryas also appears to be represented by a sudden thickening of varves. If these correlations are correct, periods of ice advances apparently were times of reduced sediment input, whereas cold intervals following the recession of ice from the basin

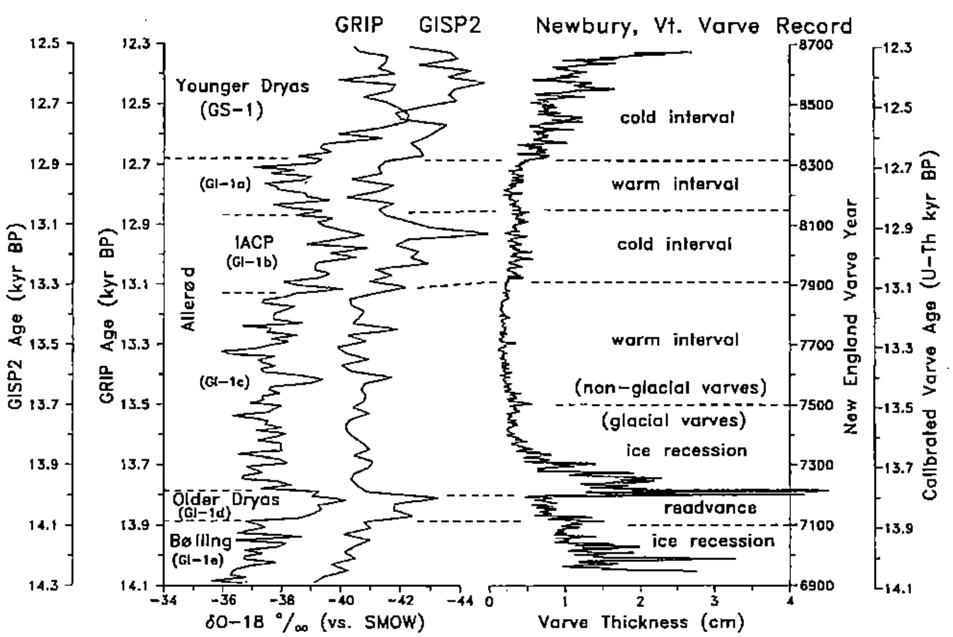

Figure 13. Comparison of oxygen isotope records from Greenland Summit ice cores (Dansgaard et al., 1998; Stuiver et al., 1995) and a 5-yr moving average of varve thickness at Newbury, Vt. Proposed event stratigraphy (Björck et al., 1998) shown in parentheses. Position of ice core time scales based on a correlation of ice core records. The Newbury record is lined up so that its calibrated time scale matches the GRIP time scale.

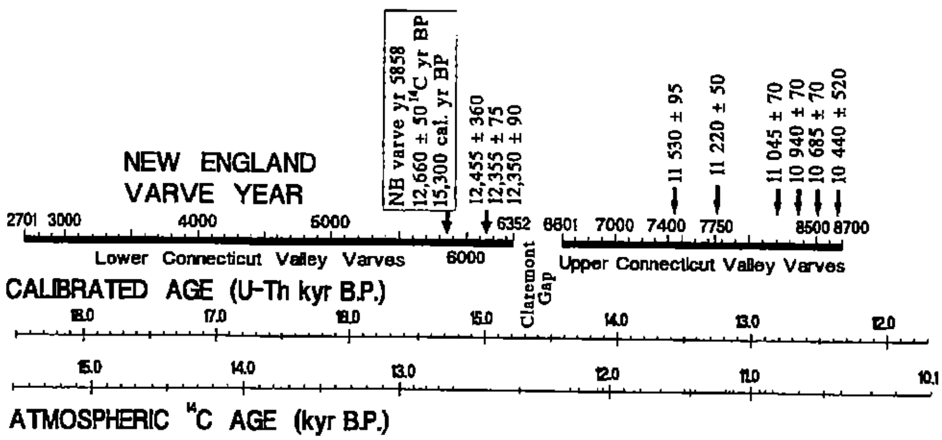

Figure 14. Atmospheric ^{14}C and calibrated (U-Th) time scales applied to the New England varve chronology (Ridge, 2003), based on the CALIB 4.3 computer program and INTCAL 98 calibration data (Stuiver and Reimer, 1993; Stuiver et al., 1998). Some individual ^{14}C ages are plotted for comparison.

were times of increased lacustrine sediment accumulation, possibly related to accelerated slope erosion.

LUNCH STOP: Quechee Gorge, Quechee, Vt.

Glacial geology of northwestern Vermont (Bierman)

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

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 Valley through a spillway near the southern end of Lake Champlain (Chapman, 1937). Glacial Lake Vermont at its lowest stage held about 240 km^3 of water, about ten times the volume of present-day Lake Champlain. Lake Vermont ended when ice-margin retreat allowed marine waters of the Champlain Sea into the isostatically-depressed Champlain Lowland. Over the next 1000 to 1500 years, isostatic rebound increased the elevation of the Richelieu River sill at a rate greater than eustatic sea-level rise, finally isolating Lake Champlain from marine water and freshening the lake by about 10,000 ^{14}C yrs. B.P. (Parent and Occhietti, 1988).

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

STOP 5-4. MT. MANSFIELD, VT.

Mt. Mansfield, at 1340 m (4393 ft) is the highest point in Vermont (Fig. 15). Mt. Mansfield's rocky summit exposes multiply-deformed schist of the Underhill Fm. (Christman and Secor, 1961), covered in places by a thin mantle of till. Although the rock has been eroded into streamlined forms, >12,000 ^{14}C yrs. 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 roughly NNW to SSE over Mt. Mansfield's summit.

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 Mts. of N.H. in the background. To the south is the spine of the Green Mts., including 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.

STOP 5-5. SMUGGLER'S NOTCH, VT.

Smuggler's Notch is a deep cleft that cuts across the main range of the Green Mts. just north of Mt. Mansfield (Fig. 15). The extreme topographic relief, recent landslide scars, and 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). Baskerville et al. (1988) described in some detail the 1983 slope failure. The rocks exposed in the cliffs above Smugglers Notch are all schist.

Rockfalls 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). We will observe the debris slide that occurred during July 13, 1983, described by Baskerville et al. (1988). The landslide began at about 7 a.m. when a large block of rock (~10.4 x 10^6 kg), 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 rockfall 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.

A debris flow deposit visible along Rt. 108 near the Cambridge/Stowe town line is one of several that occurred during the night of May 22, 1986 and are described by Lee et al. (1994). This particular flow originated in the gully extending up the east side of the valley below Spruce Peak and incorporated colluvium as well as trees and soil (~250,000 m^3 of material; Lee et al., 1994). An intense rainfall apparently precipitated this and other debris flows that evening, loosening colluvium and organic debris that had accumulated in the chute. At present, this debris flow chute is almost barren of colluvium and will take some time to once-again accumulate sufficient debris to present a hazard.

STOP 5-6. LAKE MANSFIELD AND MILLER BROOK DEPOSITS, VT.

For almost 100 years, workers have debated the presence of late Pleistocene alpine ice in the mountains of Vermont and elsewhere in New England (e.g. Goldthwaitt, 1916b; see reviews by Davis, 1999, and Waitt and Davis, 1988). Two lines of evidence have been used to argue for post-Laurentide alpine glaciation: the presence of cirque-like features and the existence of linear ridges of sediment, interpreted as moraines. Both of these landforms are well-exhibited near Lake Mansfield in the Miller Brook valley, south of Stowe, Vermont (Fig. 15).

Beginning at Nebraska Notch, Miller Brook flows through three steep-sided, bowl-shaped segments of the valley, interpreted by some as cirques (Wagner, 1971). Lake Mansfield occupies the lowest of these and is artificially dammed. Downstream from the Lake Mansfield dam, the valley contains several distinct ridges (with relief up to 30 m) lying within or adjacent to fluvial landforms and older lacustrine terraces. Wagner (1970) first identified these ridges and interpreted them as moraines produced by alpine ice retreating up Miller Brook valley. Wagner's paper stimulated much discussion and raised again the question of whether or not alpine glaciers existed in New England following retreat of the Laurentide Ice Sheet.

Based on both their map pattern and the fluvial sediments within them, we interpret these ridges to be segments of an esker, and not moraines. The poorly-sorted veneer of sediment overly-

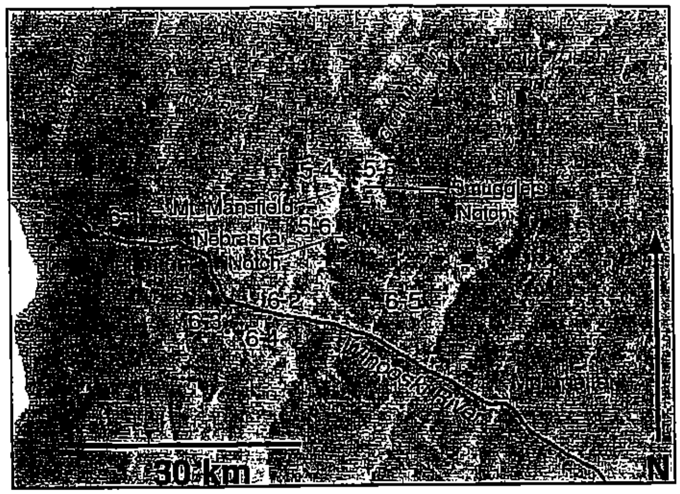

Figure 15. Digital terrain map of north-central Vermont showing the location of Stops 5-4 through 5-6 and 6-1 through 6-5.

ing the gravel likely originated from the low discharge of the stream occupying the esker tunnel towards the end of its history or from supraglacial debris dumped when the ice melted away. One ridge contains coarse diamict, including a few faceted, striated cobbles. This diamict likely melted out of the ice shortly before the ice tunnel was abandoned, so these clasts were probably never subjected to significant abrasion.

Nebraska Notch is the lowest gap through the Green Mts. between the Lamoille Valley to the north and the Winooski 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 Mts. immediately to the north and south. Once the ice elevation dropped below the elevation of Nebraska Notch, ice supply was shut off and the ice in the Miller Brook valley stagnated soon thereafter. The esker probably formed immediately before and during this stagnant ice stage. Loso et al. (1998) suggest climate could not have been cool enough in postglacial times to support a valley glacier.

DAY 6. POST-GLACIAL MASS WASTING PROCESSES, NORTHERN GREEN MTS., VT., AND SOUTHERN CONNECTICUT VALLEY (Bierman and Ridge)

STOP 6-1. TOWN LINE BROOK, WINOOSKI AND COLCHESTER, VT.

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

The valley of Town Line Brook has been widened by landsliding and deepened by fluvial incision, controlled to a great extent by the distribution of groundwater. The fine-grained sediments contain sandy interbeds along which groundwater preferentially flows. The interbeds wash out, causing small-scale slumps and toppling failures of the more cohesive, overhanging fine-grained sediments, which liquefies easily (try stamping on some failed material). The liquefaction is important because it allows failed material to be evacuated easily from the valley by rather modest stream flows. Once the fine-grained deposits have failed, the overlying, non-cohesive, and permeable deltaic sand and gravel also fail by translation and toppling. Such failures are particularly common during wet periods when the water table rises farther into the sand.

Today, only a few slides are active, but slide scars are prevalent along the watercourse. Ring counting of tree cores shows that most of the oldest trees within the currently inactive slides are about 100 years old (Baldwin et al., 1995). The age of these trees suggests that Town Line Brook hillslopes began to stabilize in the late 1800s, coincident with the reforestation of northwestern Vermont (Fig. 16). According to local residents, the major landslide complex became active within the past 20 years. Pin-line measurements suggest that the scarp has been retreating episodically over the past three years at rates of several cm to >1 m yr $^{-1}$ (Fig. 17). 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 - 15 m^3 yr^{-1}) can be calculated using valley volume (about 100,000 - 150,000 m^3) and assuming that the paleo-Winooski River delta was abandoned ~ 10,000 ^{14}C years ago when the Champlain Sea drained.

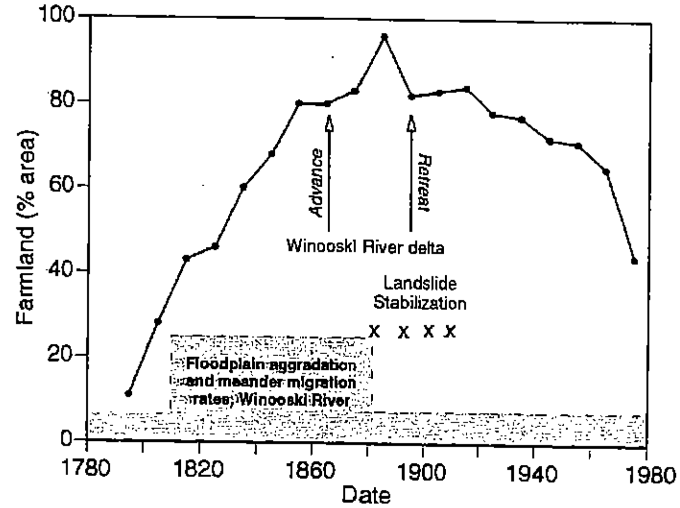

Figure 16. Historic landscape response, Chittenden County, northwestern Vermont. Open circles represent percentage farmland. Major expansion and contraction of Winooski River delta in Lake Champlain as deduced from historic maps are marked by arrows (Severson, 1991). Maximum age of trees growing in fossil 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).

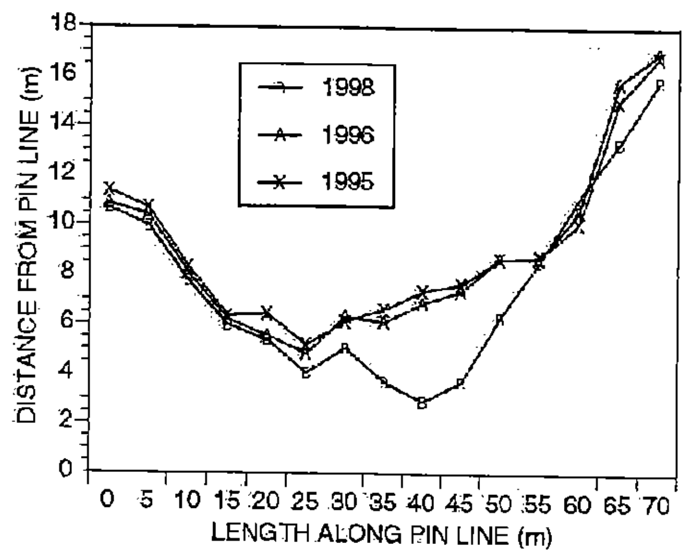

Figure 17. Pin line data for the main Town Line Brook landslide showing retreat of landslide scarp over the past three years. Data gathered by successive UVM Geohydrology classes from 1995 until 1998.

STOP 6-2. STRIATED ROCK AND ICE RETREAT, JONESVILLE, VT.

This outcrop of Underhill mica schist preserves striations and groves indicating that the last ice flowing over this outcrop moved from N70°W (Fig. 18). This flow direction is parallel to the orientation of the Winooski 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 show that striae become oriented toward regional flow directions (~NNW-SSE) at higher elevations on Camel's Hump.

In a class study, students measured lichen size as a function of underlying tombstone age in the nearby Richmond cemetery and 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. Using the Richmond calibration, 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

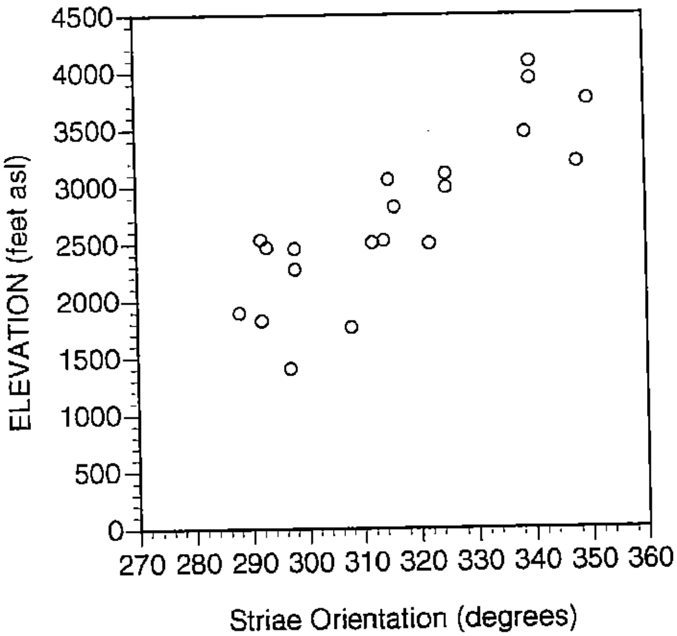

Figure 18. Orientation of striae as a function of elevation, transect from near Jonesville Rock to the summit of Camels Hump.

of the bare rock surface was most likely the result of land clearance for farming and grazing during the 1800's.

STOP 6-3. HUNTINGTON RIVER TERRACES AND ALLUVIAL FANS, HUNTINGTON, VT.

Ten terrace levels in Huntington Valley range in age from >12,500 ^{14}C years to historic. A general history of valley evolution is illustrated in Figure 19. Initially, the river flowed over and augmented the glaciolacustrine valley fill (Fig. 19). These deposits are generally a coarsening-upward sequence of rhythmically bedded silt, grading into sand (some of which may locally exhibit foresets and are therefore deltaic in origin). The fill terraces are capped by fluvial deposits. Although these fill terraces have been mapped as deltas based on their morphology (i.e., Wagner 1972), many are located far from incoming streams.

Incision and gradient changes were a response to base-level drops in the lower reaches of the Winooski Basin. The amount of incision varied along the Huntington Valley and depended on two factors: the magnitude of base-level change and the distance from the base-level change. Base-level changes began with the lowering of proglacial lakes and continued with the decline of the Champlain Sea. Each drop in water level initiated a wave of incision that propagated upstream until a bedrock knick point was encountered or until the wave of incision diffused due to lack of stream power.

Backhoe trenches and shovel pits opened on nearly every terrace here reveal three different terrace stratigraphies (Fig. 20).

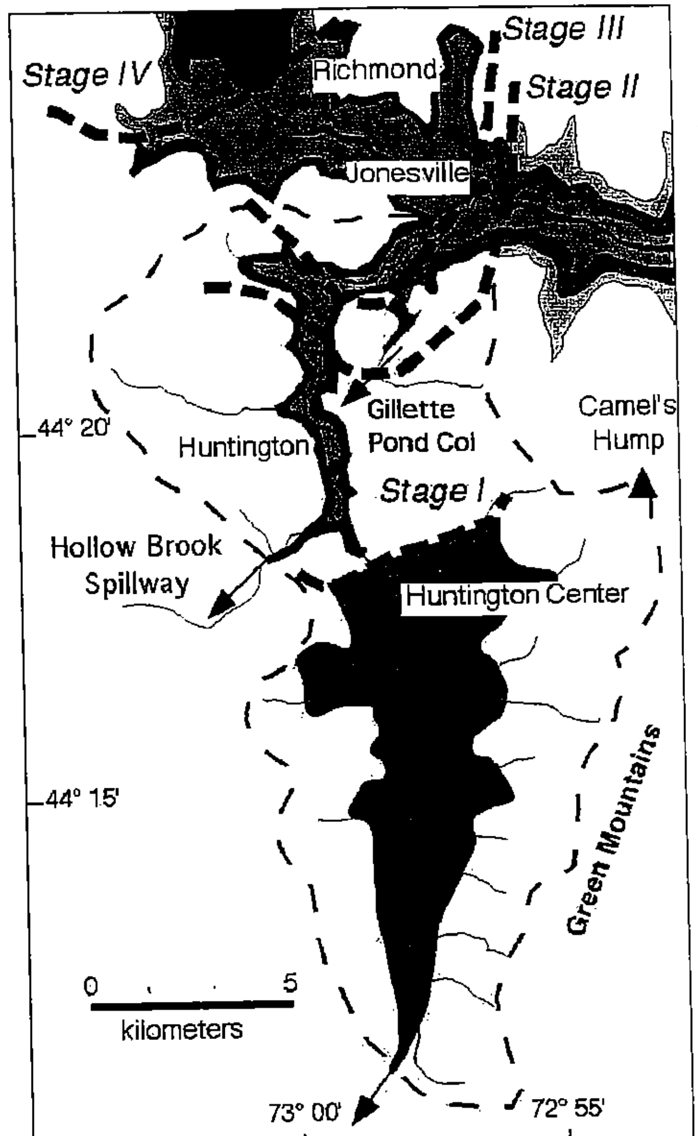

Figure 19. Map of ice margins and proglacial lake extents in the Huntington Valley of northwestern Vermont. Hollow Brook, Gillett Pond, and southern Huntington River valley thresholds are shown with arrows indicating flow directions. Ice margins are shown schematically and labeled by stage. Stage I is the earliest stage. Lake levels impounded by ice are shown in varying shades of gray. From Figure 6 of Wright et al. (1997).

Common to all terraces is the presence of a 25-cm-deep plow zone capping the fluvial deposits. The upper meter of the glaciolacustrine fill terrace, T8 (Fig. 20), is composed of cross-bedded sand overlain by silty fine sand, which suggests deposition by a meandering fluvial system. The stratigraphy of the pits and trenches from T7 and T6 (represented by the T6 log in Fig. 20) is consistent with a braided fluvial system. Overbank deposits are thin or absent on these terraces. Terrace deposits from T5 to T1 all have a basal unit of imbricated gravel, deposited in the channel, overlain by laterally-accreted sand that is capped by over-

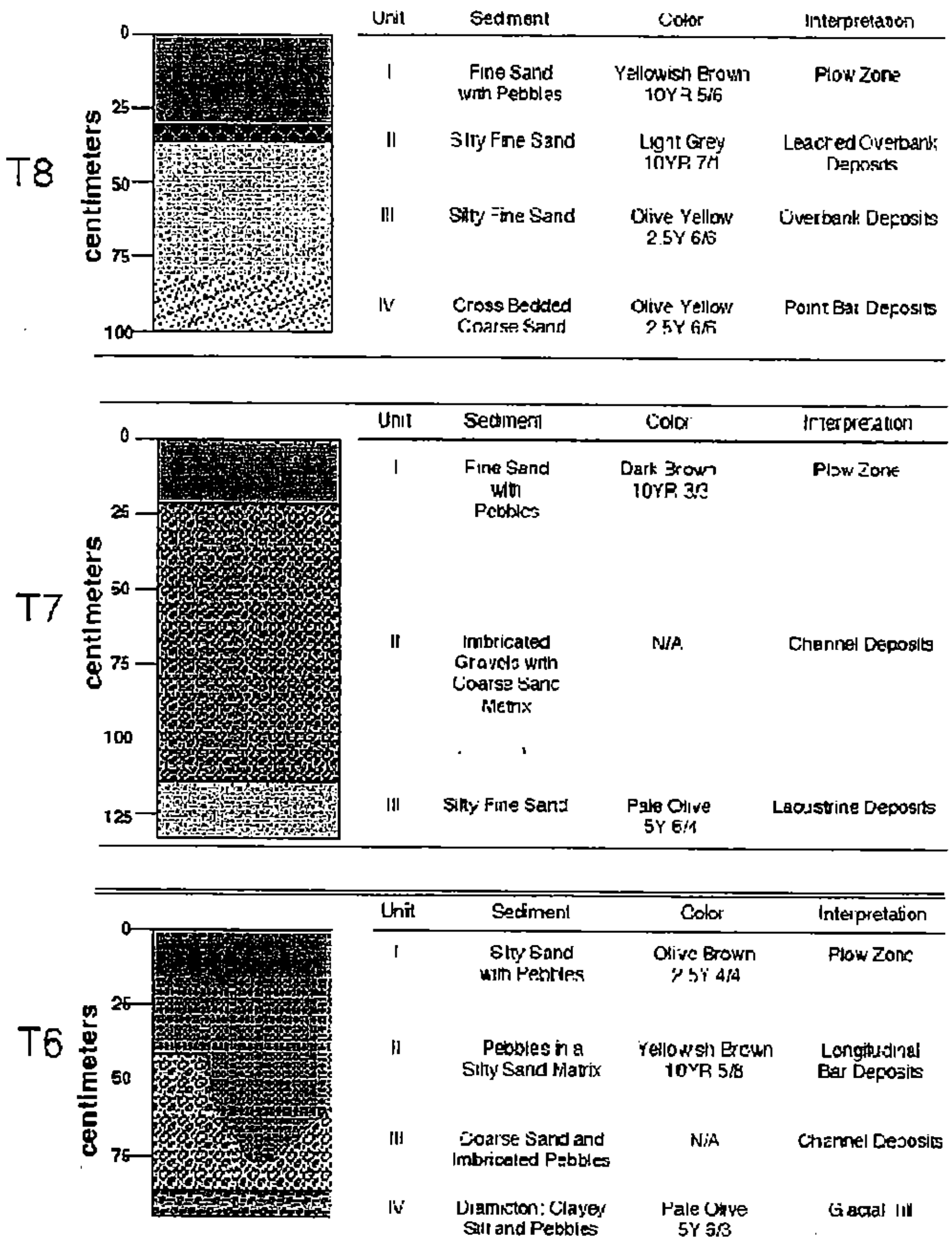

Figure 20. Examples of sedimentological difference in Huntington River terrace deposits. Oldest terraces have highest numbers. (From Whalen, 1998)

bank deposits of silty fine sand. This stratigraphy implies that the fluvial system behaved similarly to today's meandering river since the time of T5.

Alluvial fans are seldom-studied in the Northeast. We have investigated 22 alluvial fans in the Huntington River basin (Bierman et al., 1997; Zehfuss and Bierman, 1996), clustered in the northern part of the river valley where Huntington River terraces provide a platform for fan sedimentation. We have intensively studied three sites (Audubon, Moultroup, and Aldrich) and have used these sites to infer the regional history of hillslope stability during the Holocene. Recent work by Jennings et al. (2003) has extended this type of study throughout Vermont.

Most fan sediments are poorly sorted, although occasional thin (<10 cm) beds of well-sorted, clast-supported gravel occur, as well as black laminae that may represent decomposed leaf mats or concentrations of finely disseminated charcoal. Sediment deposited in the fans originates from distinct drainage basins located either on the surrounding mountainsides or on slopes created by the risers of higher Huntington River terraces. We ^{14}C-dated 14 samples of wood and charcoal from five Huntington valley alluvial fans in order to determine the timing of aggradation. These data show that two fans began to aggrade in the early Holocene (between 8530 and 8060 ^{14}C yrs. B.P., and between 7835 and 7360 ^{14}C yrs. B.P.) and two in the late Holocene (2500 and 1900 ^{14}C yrs. B.P.). One fan aggraded over 4 m during historic time (<100 ^{14}C yrs. B.P.).

Each fan contains a well-preserved, buried-soil profile, generally within the uppermost meter. But, in one fan, the soil is buried by four meters of debris, where ^{14}C ages confirm that the sediment above the buried soils was deposited coincident with historic destabilization of hillslopes. Without stability provided by root systems, the soils and sediments were more easily eroded from drainage basins, increasing sedimentation rates on the fan surfaces. Removal of vegetation interception of precipitation is also a factor. After initial deforestation, sediment continued to be deposited on the fans at a rapid rate. The recovery of forest growth on the hillsides above Vermont fans has returned some, but not complete, stability to slopes.

STOP 6-4. HUNTINGTON GORGE, HUNTINGTON CENTER, VT.

Huntington Gorge is cut through schist of the Underhill Fm. and displays well-developed potholes and plunge pools. The gorge appears to exploit joint sets trending N82°E, N75°W, and N51°W (Christman and Secor, 1961). Until recently, when the gorge formed was not known, although 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 to <11,700 ^{14}C yrs. B.P. The gorge may have been first exposed as late as 8500 ^{14}C yrs. B.P., 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. These ages lead to the important conclusion that the gorge was formed by fluvial erosion at least 1000 ^{14}C years after local deglaciation, not by the catastrophic draining of any proglacial lake.

STOP 6-5. BEN AND JERRY'S FACTORY, WATERBURY, VT., LUNCH STOP

The Ben and Jerry's factory is currently the most popular tourist attraction in Vermont. The factory is located at an elevation of about 180 m, an area that would have been on the bottom of high-level lakes in the Winooski Valley, but near the shoreline as the ice sheet retreated, establishing glacial Lake Vermont (Figs. 1, 12). By Champlain Sea time, the factory would have stood on dry land.

STOP 6-6. CANOE BROOK SECTION, DUMMERSTON, VT. (Ridge)

The Canoe Brook section has been important in establishing the chronology of glacial Lake Hitchcock and calibration of the lower Connecticut varves (Figs. 1, 12). Ridge and Larsen (1990) measured the varves and matched the sequence to varves 5685-6229 of the lower Connecticut varves. Thick ice-proximal varves at the base of the section also give an estimate of the age of deglaciation in southern Vt. More important, several terrestrial plant fossils recovered from high in the section (varve 6150) yielded the first ^{14}C ages (average 12.3-12.5 ka ^{14}C yrs. B.P. or 14.3-14.8 cal. ka) from glacial Lake Hitchcock sediment. This allowed the first calibration and application of both ^{14}C and calibrated (U-Th) time scales to the lower Connecticut varves (Ridge and Toll, 1999; Ridge et al. 1996, 1999). Unfortunately, these ^{14}C ages are from the later stages of a plateau in the ^{14}C time scale (12.6-12.4 ka ^{14}C yrs. B.P.) when about 200 ^{14}C years span about 900 calibrated (U-Th) years. This situation led to uncertainty in the calibration of the varve chronology. Recently, another ^{14}C age at Canoe Brook from lower in the section that pre-dates the ^{14}C plateau (varve 5858, 12,660 $\pm$ 50 ^{14}C yrs. B.P. or 15,300 +227/-105 cal. yrs.) has provided a more precise calibration of the lower Connecticut varves (Ridge, 2003; Fig. 14).

DAY 7. GLACIAL LAKE HITCHCOCK, CONNECTICUT VALLEY, WESTERN MASS. (Brigham-Grette and Rittenour)

The Connecticut Valley in western Mass. contains a rich history of Wisconsin deglaciation and the drainage history of glacial Lake Hitchcock (Fig. 21). Studies in this region go back more than 150 years, but most surficial geologic maps for the areas we will examine were completed by Jahns, who was instrumental in developing the morphosequence concept (Jahns and Willard, 1942). The text that follows was compiled by a number of individuals for a recent field conference (Brigham-Grette et al., 2000). Earlier work of Koteff and Larsen (1989) suggested that glacial Lake Hitchcock drained at once from end to end due to erosion of the sediment dam at Rocky Hill, Conn. More recent work demonstrates that drainage of individual basins was sequential within the lake system, a process that took several thousand years as the Laurentide Ice Sheet retreated (Ridge et al., 1999; K. Curran and T.M. Rittenour, unpublished data). Direct dating of the varve sequence in the lake (Ridge et al., 1999) and sedimentological study of the varves indicate that they record an El Nino / Southern Oscillation-like (ENSO) signal for the period from about 17,000 to 13,000 cal. yrs. B.P. (Rittenour et al., 2000). This record is the first ever to demonstrate a teleconnection to western-Pacific, ENSO-like activity during the LGM.

STOP 7-1. BARTONS COVE KNICKPOINT AND GLACIAL LAKE HITCHCOCK DRAINAGE, TURNERS FALLS, GILL, MASS.

During the stable phase of glacial Lake Hitchcock, the Montague delta (fed by the Millers River) extended completely across the lake basin (Figs. 21, 22), evident from the delta foreset directions along the west side of Canada Hill and north of the Mineral Hills. The Montague delta separates the narrow, shallow-lake

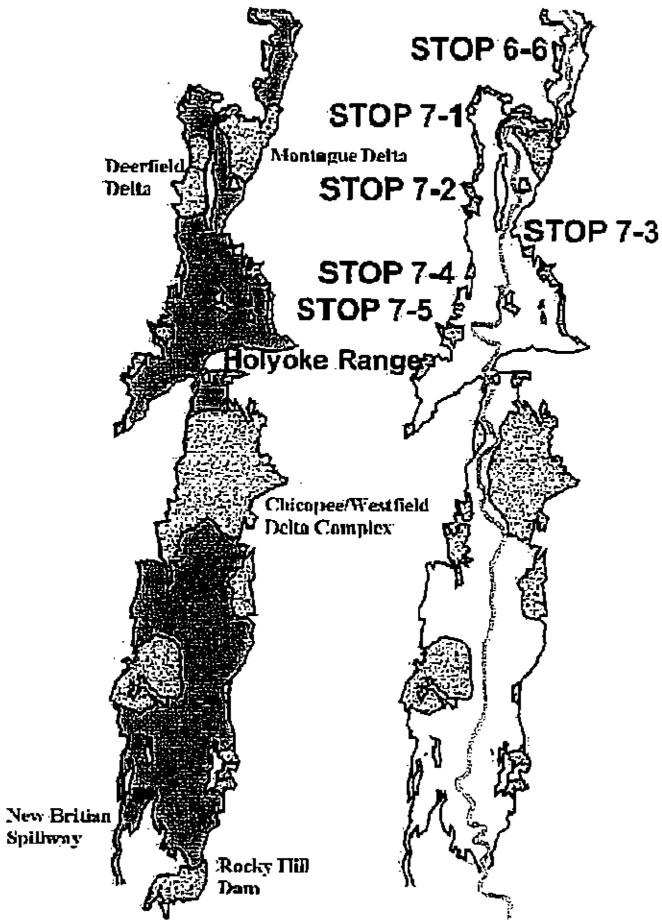

Figure 21. Schematic map showing stops for DAY 7. The sketch outlines glacial Lake Hitchcock in Mass. and Conn.; left sketch is lake filled; right sketch shows course of the modern Connecticut River. Stippled areas are prominent deltas graded to the highest "stable" lake level.

basin to the north (in the Northfield, Mass., area) from the deep, broad, Hadley basin to the south. The Hadley basin was separated from the remaining portions of glacial Lake Hitchcock to the south by the large Chicopee-Westfield delta complex, which also grew to extend across the lake basin.

Lake levels north of the Holyoke Range were controlled by incision into the Rocky Hill, Conn., sediment dam and subsequent entrenchment of the proto-Connecticut River into the Chicopee-Westfield Delta complex. The first lake-level drop in the Hadley basin is known as Lake Hadley (Fig. 23) and is recorded in topset / foreset contacts in the southernmost stretch of the Montague delta, 3-5 m below the stable lake level (Emerson, 1898; Jahns and Willard, 1942; Jahns and Lattman, 1962; Jahns, 1967; Koteff and Larsen, 1989). This drop in lake level may correlate with the terrace cut into the Rocky Hill sediment dam area at 2-5 m below the stable lake level and other deltas south of the Holyoke Range built at this level (Koteff and others, 1988; Koteff and Larsen, 1989; Stone and Ashley, 1992).

During initial lake drainage, a proto-Connecticut River channel was developed on the west side of Canada Hill (Fig. 22). This channel, called the White Ash Swamp channel, has a base at 84 m (270 ft) (~20 m below the stable lake level here) and is the oldest surface occupied by the Connecticut River during incision into the Montague delta. The White Ash Swamp channel was abandoned during further river incision and development of a more favorable route, similar to that of the modern Connecticut River, on the east side of the Canada Hill bedrock ridge.

Complete entrenchment through the Chicopee-Westfield delta complex eventually drained the Hadley basin. At this point, early strath terraces, cut by the flow of water across the lake bottom, were formed. Within a short period of time, channel flow was initiated in the proto-Connecticut River, and these early subtle terraces were abandoned. The exposed, lake-bottom sand quickly became subject to eolian transport, generating dunes and sand sheets. To the north, the drop in base level from the drainage of Lake Hadley caused the Connecticut River to entrench deeper into the Montague delta (Fig. 23). During incision the river became perched on a bedrock ledge near Turners Falls, Mass. (Fig. 22). Erosion continued south of this knick point but was suspended to the north. This situation created two waterfalls over the sandstone ridge, with associated plunge pools at the base (Fig. 22). Prior to the flooding of Barton's Cove by construction of the Turners Falls dam more than 100 years ago, these plunge pools were known as Lily Ponds and so the bedrock ridge has become known as the Lily Pond barrier (Emerson, 1898). Deep channels were cut into the barrier and prominent terraces formed upstream as the Connecticut River flowed over this knick point. Geomorphic and archeological evidence suggest this knick point may have been active for thousands of years, possibly having been abandoned before 10,200 cal. yrs. B.P. (9,000 ^{14}C yrs. B.P.) (Curran, 1999).

STOP 7-2. CONNECTICUT VALLEY VARVES, RIVER ROAD SITE, SUNDERLAND, MASS.

The River Road varve site, located just north of Sunderland (Fig. 21), lies on the west side of the Connecticut River and on the eastern flank of the Pocumtuck Range (Mt. Sugarloaf). The site lies midway between the Montague Plain (topset/foreset contact = 102.7 m) and Long Plain delta (89.9 m) and is one of the most studied exposures in this part of the valley. The inferred lake level at this site is 95.1 m. Two small slumps along this small creek expose over 200 years of sedimentation in glacial Lake Hitchcock. The two exposures have been correlated to each other (Levy, 1998) and to the Antev's master curve (varve #s 4980 to 5200) using patterns of varve thickness (Thomas, 1984). Using Ridge et al.'s (1999) calibrated varve curve, the age range for the site is inferred to be ca. 13,988 - 13,146 cal.yrs. B.P.

Carbonate concretions (40-45 wt. %) found at the River Road exposures range in size from 1 to 5 cm (long axis) and their morphologies appear to be closely related to the micro-stratigraphy of the host sediment (Levy, 1998). Based on carbon isotope data, the source of the concretion carbonate is

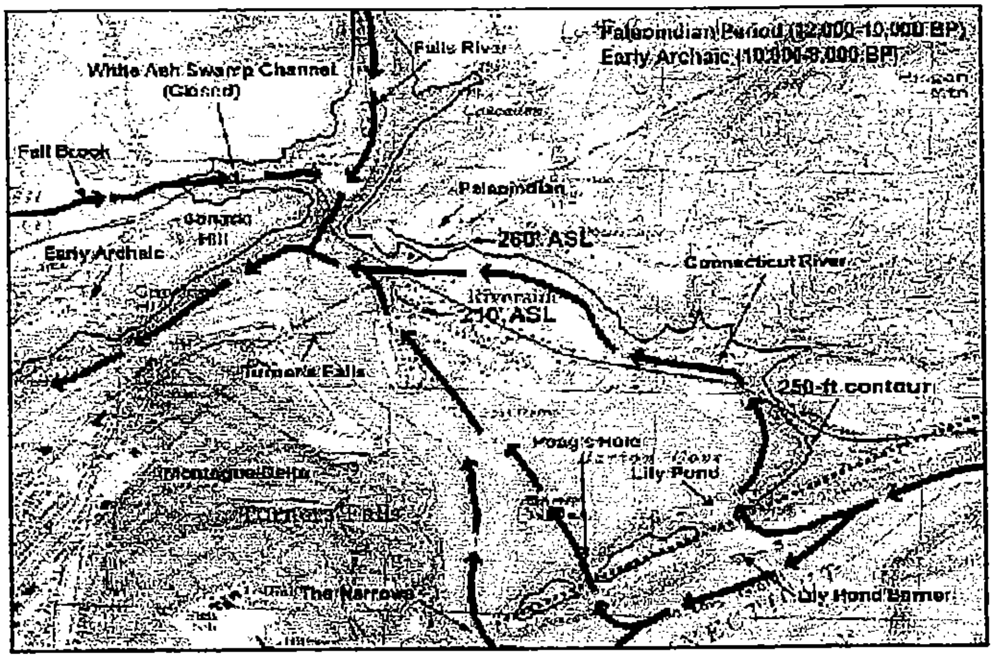

Figure 22. Part of Greenfield, Mass., 1:25,000 quadrangle, patterned area represents the once-continuous surface of the Montague delta connected by arrows. White Ash Swamp channel and Lily Pond barrier are labeled.

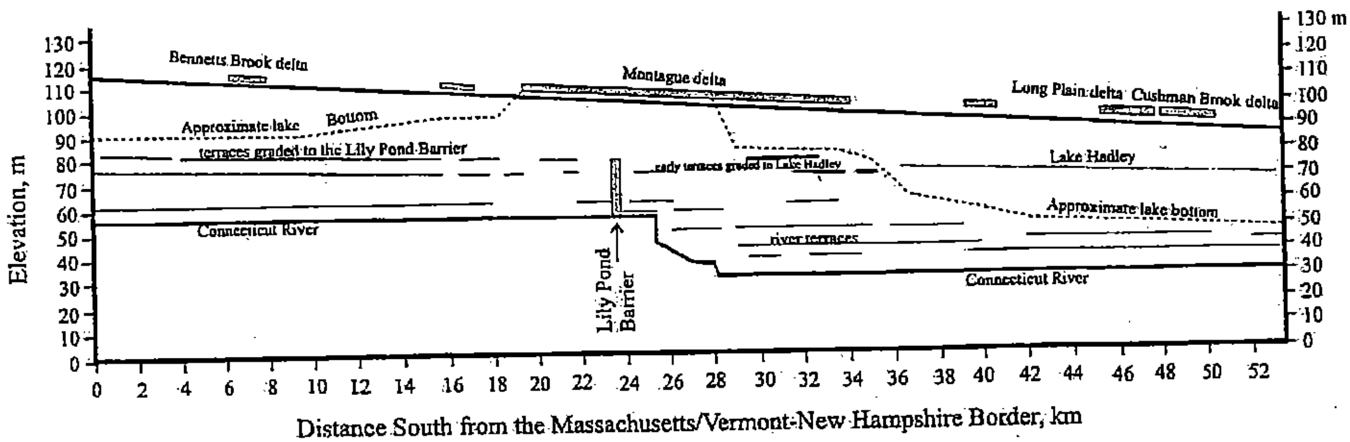

Figure 23. Longitudinal profile of highest Hitchcock terraces in the Hadley and Northfield basins, Mass. (from Rittenour, 1999). Dotted line is the ancient lake bottom and thick solid line the elevation of the modern river system. The Lilly Pond barrier at Barton's Cove in Turner's Falls, Mass., is shown schematically as a small vertical wall.

thought to be carbon originating from terrestrial organic matter deposited with the lake sediment. Individual concretion bands have been AMS ^{14}C-dated to evaluate their utility for reliable age control. Concretion centers date up to 2,500 years too old and the outer concretion samples too young (relative to the inferred age of the site), questioning the reliability of this approach. Recent work by Bosiljka Glumac (Smith College) indicates oxygen isotope ratios of –8.5 to –13.0 %δ (SMOW), values consistent with interglacial meteoric (local) water.

STOP 7-3. SUNDERLAND DELTA, MORPHOSEQUENCES AND ECHO DUNES, SUNDERLAND, MASS.

The Sunderland delta (or Long Plain delta, Fig. 24) is a classic example of the ice-contact / meteoric delta, graded to the stable level of glacial Lake Hitchcock. The topset / foreset contact occurs here at 90 m (295 ft). Once the ice margin had retreated northward and bifurcated around Mt. Toby, rain and meltwater continued to prograde the delta to lake level.

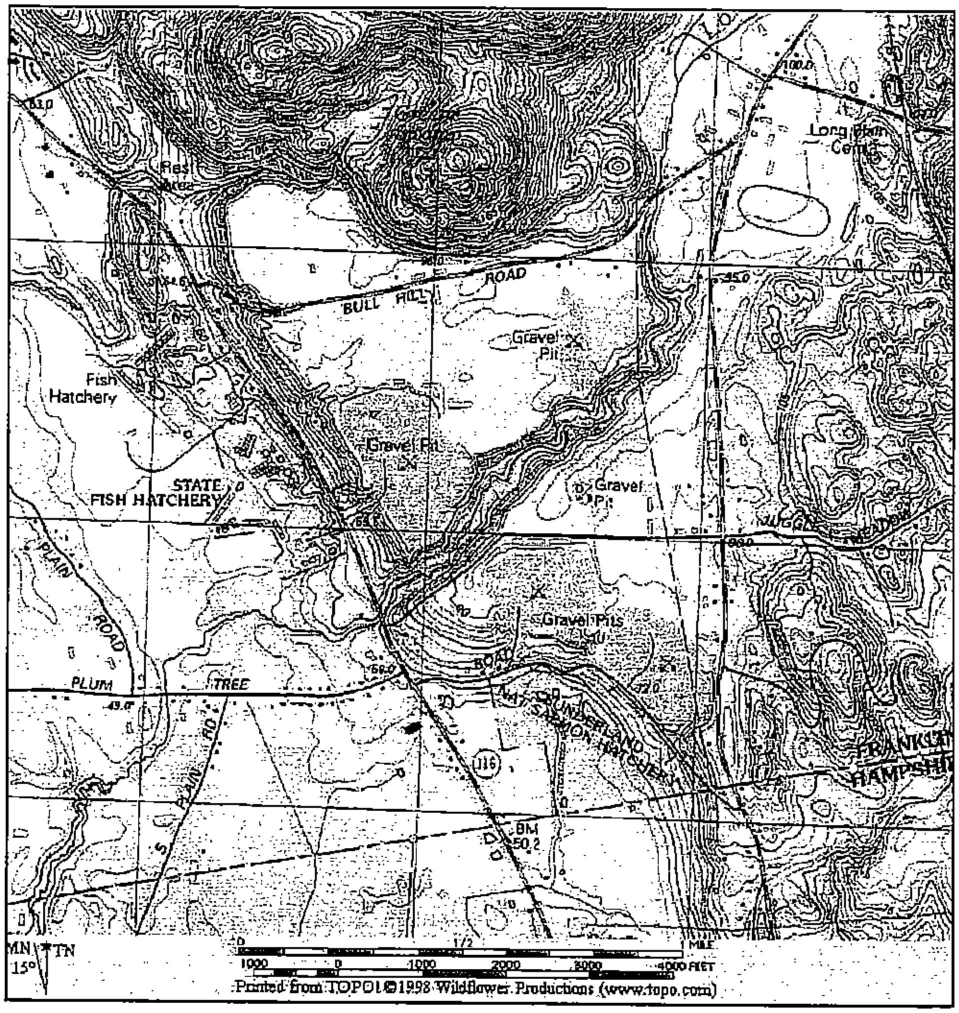

Figure 24. Part of Mt. Toby, Mass., 1:25,000 quadrangle, showing the Sunderland delta. Note the narrow canyon cut by the Long Plain River as glacial Lake Hitchcock drained and the strategic location of the fish hatcheries tapping freshwater springs from the base of the delta. Route 63 cuts through the echo dune at the base of the delta.

Following drainage of glacial Lake Hitchcock from the Hadley basin, Long Plain Brook quickly adjusted to a lower base level by incising a small canyon through the delta (Fig. 24). This downcutting produced an alluvial fan out over the floor of the former lake. Today, groundwater maintains a continuous flow through the delta, producing natural springs from sand and gravel beds confined by lake clay and silt at the foot of the delta slope. The State of Massachusetts maintains two fish hatcheries for trout at the base of the delta.

Periglacial conditions persisted in the valley for some time following lake drainage. Here, in front of the Sunderland delta, winds blowing across the exposed sediments of glacial Lake Hitchcock picked up sediments and created an echo dune. Such dunes are formed due to air turbulence as wind rises over an obstacle, commonly a steep bedrock outcrop, drumlin, or delta front on the eastern side of the lake basin.

LUNCH STOP: Mt. Sugarloaf State Park, Sunderland, Mass. (alternative site Mass. State Fish Hatchery or Yankee Candle Company picnic area)

STOP 7-4. UNIVERSITY OF MASSACHUSETTS-AMHERST NATIONAL GEOTECHNICAL EXPERIMENTATION SITE; ENSO-LIKE SIGNALS IN GLACIAL LAKE HITCHCOCK VARVES, AMHERST, MA

In the fall of 1997, two 10-cm-diameter cores were drilled at this site (Fig. 21), 1.5 km from the shore of glacial Lake Hitchcock, on a 43 m, lake-bottom surface in the broad, deep Hadley basin. In this area, the lake was about 20 km wide and the stable lake level was at 90 m. Water depth at the core site was 77.5 m at the initiation of varve deposition. Two cores were drilled to study the classic glacial varves in an area of the valley that lacked natural exposures and represented a large geographic gap in Antev's original varve record. One core was drilled to bedrock, and a second shorter core was taken to ensure complete recovery of the upper portion of the varve sequence (7.6 m). The U. Mass. core covers varves 4638-6027, capturing at one site a total of 1,389 varves, or nearly one-third of the entire lake history. The 32-m-long core drilled to bedrock provided a continuous sequence of varves ranging from thick (55 cm) ice-proximal varves at the

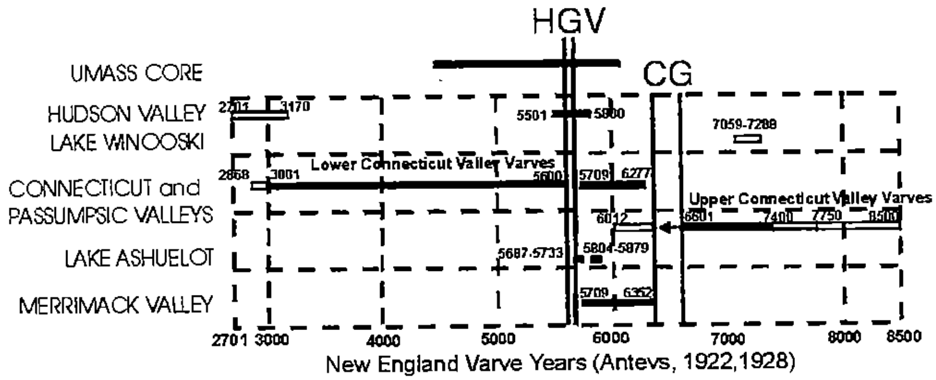

Figure 25. Time span in varve years (younger to the right) of overlapping sequences of the NE varve chronology (NE 2701 – 8500) plus the UMass core (4638 – 6027). Hudson Valley gap (corrected by Rittenour, 1999) and Claremont gap (corrected by Ridge et al., 1999)

base, through varves of average thickness for the region (1-4 cm), and into thin varves with relatively coarse-grained summer layers and thinner winter clay layers.

A ^{14}C age from small plant fragments picked from varves in a contorted zone between varves 5761-5768 produced a $\delta\,^{13}C$-corrected AMS age of 12,370 ± 120 ^{14}C yrs. B.P. (14.3 +1.2/-0.4 cal. kyr B.P.). AMS ages from the top and bottom of the core indicate that it records varve deposition from 15.4 to 14.0 cal. kyr B.P. (12.8 to 12.0 ka ^{14}C yrs. B.P., coincident with a large ^{14}C plateau at this time where ~1400 cal. yrs. equals only 800 ^{14}C years). The varves were correlated to normal curves from glacial Lake Hitchcock (datasets MASS 4-13, NH 13, VT 14-15, and VT-NH 16-17; Antevs, 1922) and the Canoe Brook section of Ridge and Larsen (1990). A 300-year section of the U. Mass. core was also matched with a normal curve from the Hudson Valley (dataset NY 13-14; Antevs, 1922) and used to bridge a critical gap in the Connecticut Valley sequence (Rittenour, 1999). This gap and the Claremont Gap in Vt. filled by Ridge et al. (1999) are the only two major adjustments that have been made to the Antevs' varve chronology. Other varve thickness measurements and correlations in the over 4000-year varve chronology have otherwise withstood vigorous testing (Fig. 25).

The chronology and varve counting from this site suggest that glacial Lake Hitchcock completely drained in the Hadley basin sometime after varve 6027 was deposited at the top of the core) at 14.0 ka cal. yrs. B.P. (12.0 ka ^{14}C yrs. B.P.). Counting to the base of the section, Laurentide ice had retreated from the Amherst, Mass. area as the first, ice-proximal varve was deposited (varve 4638) at 15.4 ka cal. B.P. (12.8 ka ^{14}C yrs. B.P.).

Spectral analyses were conducted on the New England varve chronology (varves 2868-6900) to test for climate signals in the varve-thickness record (Rittenour et al., 2000). Statistically significant climate signals occur at multi-decadal (>40 year) timescales, within the conventional 2.5-5 year and 7-9 year cycles associated with ENSO. The 22-year period signal also seen here has the same frequency as the Hale solar magnetic cycle, which has been correlated with modern Northern Hemisphere temperature variations. The absence of the lower-frequency component of variability during the latter part of the interval studied here (i.e., the latest Pleistocene) is consistent with the theoretical prediction of fewer large ENSO events in the early Holocene. Rittenour et al.'s (2000) results suggest that the ENSO system was operational during the late Pleistocene, when the Laurentide Ice Sheet was near its maximum extent and climatic boundary conditions were different than today.

STOP 7-5. GEOCHRONOLOGY OF DUNES ON THE FLOOR OF GLACIAL LAKE HITCHCOCK, HADLEY, MASS.

A variety of fossil dunes can be found throughout the Connecticut Valley and most are thought to have formed under cold, dry conditions just after the drainage of glacial Lake Hitchcock. Dunes are found on a variety of surfaces but of critical importance are those that formed on abandoned deltas and the floor of glacial Lake Hitchcock (Rittenour, 1999). Exposed at this site are sediments from a sand dune that formed on exposed lake-bottom sediments in the Hadley basin (Fig. 21). It is one of many roughly north-south-trending, transverse dunes and is composed of shallowly-dipping beds that mimic the gentle lee slope of the dune.

Optically-stimulated luminescence (OSL) ages were obtained from parabolic and transverse sand dunes located on a number of genetic surfaces within the Connecticut Valley, Mass. (Rittenour, 1999; OSL analysis by Steve Forman, Univ. Illinois, Chicago). The ages from these dunes indicate that most formed between 14.0 ± 1.0 and 14.4 ± 1.0 ka cal. yrs. B.P. (12.0 to 12.4 ka ^{14}C yrs. B.P). An OSL age on one dune suggests that the Hadley basin completely drained prior to the deposition of the sand dune at 14.3 ± 1.6 ka cal. yrs. B.P. (12.3 ka ^{14}C yrs. B.P.). Also, ^{14}C ages from the U. Mass. core indicate that varve deposition continued within the Hadley basin until 14.0 ka cal. yrs. B.P. (12.0 ka ^{14}C yrs. B.P.). These ages constrain the complete drainage of glacial Lake Hitchcock in Massachusetts to between 14.0 and 14.3 ka cal. yrs. B.P.

End of field trip! Proceed to Logan International Airport, Boston, Mass.

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