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EMERITUS PROFESSOR OF GEOLOGY

JOSEPH H. HARTSHORN
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GLACIAL LAKE ALBANY AND ITS SUCCESSORS IN THE HUDSON LOWLANDS

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INTRODUCTION

Glacial Lake Albany was a northward-expanding proglacial lake that developed in the Hudson Lowlands as the Hudson-Champlain glacial lobe retreated (Figs. 1a and b). The receding ice margin formed the northern shore of Lake Albany from Long Island to the latitude of the Batten Kill. The lake extended from Glens Falls, NY, to Long Island, NY, and included Glacial Lake Hudson in the lower Hudson Valley. The water level of the lake fell abruptly to the Lake Quaker Springs stage as the glacier retreated into the Champlain Valley. Lake Quaker Springs was an extension of Glacial Lake Vermont. Lower stages of Lake Vermont include Lakes Coveville and Fort Ann. These water levels can be traced into the Hudson Valley, where they become more "fluvial" and less "lacustrine."

Lakes Albany and Quaker Springs are recorded by sand and silt terraces, beaches, and deltas throughout the Hudson Lowlands (Woodworth, 1905; Cadwell and Dineen, 1987). The elevations of these terraces and deltas have been tilted up to the north by glacial-isostatic rebound (Woodworth, 1905; LaFleur, 1965a; Connally and Sirkin, 1986; DeSimone and LaFleur, 1986). Lake Albany stage deltas disappear north of Saratoga Lake, and Lake Quaker Springs features can be traced north into the Champlain Valley (DeSimone and LaFleur, 1986). The terraces associated with Lakes Coveville and Fort Ann are not tilted to the same degree as the older lake features (Woodworth, 1905; Connally and Sirkin, 1986; DeSimone and LaFleur, 1986). The lower lake stages are recorded by fluvially-eroded terraces, spillways, catastrophic flood channels, small deltas, and infrequent beaches.

Lake Albany received large quantities of glacial meltwater from the retreating ice margin. Large tributary basins, particularly north of Albany, fed vast amounts of water into the later lake stages. Glacial Lake Iroquois, a large proglacial lake in the Ontario Lowlands, drained into the Mohawk Valley via the Rome-Little Falls outlet (Fig. 1c; Muller and Prest, 1985). The Iroquois waters entered Lake Albany at Schenectady, NY, where they built a large, sandy delta (LaFleur, 1979; Dineen and Rogers, 1979). Catastrophic floods from Lake Iroquois carved large, deep channels that were graded to the Quaker Springs through Coveville stages of Lake Albany (Stoller, 1920; LaFleur, 1965b; Hanson, 1977). Lake Iroquois and its successor, Lake Frontenac, eventually drained through the Covey Hill Spillway into Lake Vermont when the Ontario Lobe retreated north of the Adirondack Mountains (Chapman, 1937; LaFleur, 1965a; Muller and Prest, 1985).
Figure 1c. GLACIAL LAKES

OUTLETS:
1. GRAND GORGE
2. FRANKLIN/ CATSKILL
3. DELANSON
4. KAYADEROSSEAS
5. FORT ANN
As the glacier retreated into the Champlain Valley, Lake Vermont formed between the Hudson-Champlain divide and the ice front (Chapman, 1937). Waters escaping from Lake Vermont carved a series of channels near Fort Ann (DeSimone and LaFleur, 1985, 1986). Lakes Coveville and Fort Ann are stages of Lake Vermont that were broad rivers in the Hudson Valley (LaFleur, 1965a).

The Lake Albany-Lake Vermont sequence of lakes ended when the ice front receded from the Covey Hill spillway into the St. Lawrence Lowlands (LaSalle, 1966). This event allowed the Champlain Sea to invade the St. Lawrence and Champlain Lowlands, and cut off the supply of glacial meltwater to the Mohawk, Champlain, and Hudson Valleys (Clark and Karrow, 1984; Muller and Prest, 1985).

GEOGRAPHY

Glacial Lake Albany sediments lie in the Hudson-Champlain Lowlands of eastern New York (Figs. 1a and c; Cadwell and Dineen, 1987). These sediments form extensive clay terraces that border the Hudson River from Newburgh, NY, north to Glens Falls, NY (Merrill, 1890, 1891; Ries, 1890; Upham, 1903; Peet, 1904; Woodworth, 1905). Similar clay terraces lie along the Hudson River south of the Hudson Highlands from Peekskill, NY, to Staten Island, NY (Merrill, 1890, Ries, 1890; Jones, 1899; Woodworth, 1905; Reeds, 1927). The riverside terraces are underlain by till from Newburgh to Peekskill (Upham, 1903; Peet, 1904; Woodworth, 1905). The clay terraces grade into higher sand and gravel terraces away from the river.

The Hudson Lowlands are underlain by Lower Paleozoic shale and sandstone (Fisher and others, 1970). The New England Uplands, to the east of the lowlands (Fig. 1a), are underlain by metamorphosed Precambrian through Lower Paleozoic igneous and sedimentary rocks. The Catskill Mountains and Appalachian Plateau lie west of the lowlands. They are underlain by slightly deformed sediments of Silurian to Upper Devonian age. The Triassic Lowlands south of the Hudson Highlands are underlain by slightly deformed sedimentary and igneous rocks. The lowlands are bordered to the northwest by the Precambrian metasedimentary and metaigneous rocks of the Adirondack Highlands.

The Hudson-Champlain Lowlands were drained by a trellis drainage network in preglacial time (Fig. 2a; Dineen and others, 1983; Dineen, 1987). This network had been rejuvenated in late preglacial time. Deep, north-south trending, parallel gorges and sharp breaks in the stream gradients were formed in late preglacial or early glacial time, apparently in response to lower sea level. The Colonie, Battenkill-Hudson, and Mohawk channels were the trunk streams at this time (Dineen and Rogers, 1979). Only the Battenkill-Hudson channel can be traced south of Ravena (Fig. 2a; Dineen, 1987).
The Hudson Lowlands are connected to the Ontario Lowlands by the Mohawk Valley, and to the St. Lawrence Lowlands by the Champlain Valley (Fig. 1a, 1c, 2a). The ranges of the New England Uplands separate the Hudson and Connecticut Lowlands.

These large-scale geographic features controlled the flow of the Wisconsinan Glacier through eastern New York State (Fig. 1b; Hughes and others, 1985). The Adirondack Highlands split the flow of the eastern section of the Laurentide Ice Sheet into the St. Lawrence, Ontario, and Hudson-Champlain Lobes. Ice flow through the Adirondacks was weak, and the Ontario and St. Lawrence Lobes were able to push the Oneida sublobe into the western Mohawk Valley, while the Hudson-Champlain Lobe pushed the Mohawk sublobe into the eastern Mohawk Valley.

The Hudson-Champlain Lobe reached the western end of Long Island. The Connecticut Lobe flowed south to Long Island via the Connecticut Lowlands, and the Narragansett Lobe reached eastern Long Island via Narragansett Bay (Fig. 1b). The relative ages of radiocarbon-dated bogs, ice-flow models, and cross-cutting moraines suggest that Hudson-Champlain Lobe began to retreat before the others (Connally and Sirkin, 1973; Hughes and others, 1985; Sirkin, 1986; Dineen, 1986a).

Large glacial lakes occupied the various lowlands as the Wisconsinan glacier retreated (Figs. 1c and 2b). The detailed history of these lakes is still being studied, although their relative stratigraphy is becoming clear. The lakes became younger from south to north and east to west. They communicated with each other as the retreating ice front exposed lower interconnecting spillways and cols (Figs. 1c and 2b, Table 1).

SEDIMENTOLOGY

The earliest mappers in the Hudson Valley recognised the complex nature of the Lake Albany deposits. Eights (1852) described widespread Hudson Valley clays draped over boulder clay (till), which covered striated bedrock. Ries (1890), Jones (1899); Peet (1904), and Woodworth (1905) all noted that the clays were underlain by striated bedrock, till, or sand and gravel. Ries (1890) described localities where the sand and gravel was faulted (especially at Hudson and Coeymans Landing— see Stop 5). Peet (1904) noted that the cross-bedding in these basal gravels usually dipped away from the river and that the tops of these gravel deposits tended to be quite irregular. Woodworth (1905) agreed with the interpretation of Ries (1890) and Peet (1904) that the basal sand and gravel masses were kames.

Eights (1852) divided the overlying clays into two units— a lower blue clay with seams of fine sand and an upper weathered, brown, concretionary clay that was locally contorted with north-south grooves and "injected" with lenses of sand and gravel. Ries
Figure 2b. GLACIAL LAKES
<table>
<thead>
<tr>
<th>AGE IN YEARS</th>
<th>ONTARIO LOWLANDS</th>
<th>MOHAWK VALLEY</th>
<th>HACKENSACK LOWLANDS</th>
<th>HUDSON LOWLANDS</th>
<th>CHAMPLAIN LOWLANDS</th>
<th>ST. LAWRENCE LOWLANDS</th>
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<td>10K</td>
<td>Early Lake Ontario&lt;sup&gt;1&lt;/sup&gt;</td>
<td></td>
<td></td>
<td>Estuary&lt;sup&gt;11&lt;/sup&gt;</td>
<td>Champlain&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Champlain&lt;sup&gt;1&lt;/sup&gt; Sea</td>
</tr>
<tr>
<td>12K</td>
<td>Lake Frontenac&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Lake Iroquois&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Catastrophic Floods&lt;sup&gt;4&lt;/sup&gt;</td>
<td>Hudson River</td>
<td>Fort Ann&lt;sup&gt;3&lt;/sup&gt;</td>
<td>Fort Ann</td>
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<td></td>
<td>Lake Frontenac&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Lake Iroquois&lt;sup&gt;1&lt;/sup&gt;</td>
<td>Catastrophic Floods&lt;sup&gt;2&lt;/sup&gt;</td>
<td>Free Drainage</td>
<td>Lake Coveville</td>
<td>Coveville</td>
</tr>
<tr>
<td>14K</td>
<td>ICE</td>
<td>Lake Amsterdam</td>
<td>&quot;Fluvial Sand and Gravel&quot;&lt;sup&gt;6&lt;/sup&gt;</td>
<td>Lake Quaker Springs</td>
<td>MDM&lt;sup&gt;4&lt;/sup&gt;</td>
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<td>Readvance&lt;sup&gt;5&lt;/sup&gt;</td>
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<td>I</td>
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<td>Lake Albany</td>
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<td>SDM&lt;sup&gt;7&lt;/sup&gt;</td>
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<tr>
<td>18K</td>
<td>E</td>
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<td>Lakes Hudson&lt;sup&gt;12&lt;/sup&gt; &amp; Flushing</td>
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1 Muller and Prest, 1965  
2 Muller, Franz, Ridge, 1986  
3 DeSimone & LaFleur, 1986  
4 Dineen, 1986  
5 LaFleur, 1979  
6 Lovegreen, 1974  
7 Connally & Sirkin, 1986  
8 Dineen & Rogers, 1979  
9 LaFleur, 1965a  
10 Clark & Karrow, 1984  
11 Weiss, 1974  
12 Reeds, 1927

WM = Wallkill Moraine  
RHM = Red Hook Moraine  
MDM = Meadowdale Moraine  
SDM = Shenandoah Moraine
(1890) described the clays as blue and yellow marly clays, yellow on top, with lenses or layers of "wavy" silt loam or sand. He also noted concretions in the upper clay. Peet (1904) emphasized the laminated nature of the clay, with "fatty" laminae alternating with sandy laminae. He also noted that the upper clays were yellow while the lower were blue, and joints and contorted bedding were common. Woodworth (1905) noticed that the clays south of the Hudson Highlands were often contorted, and frequently contained lenses of till. Ries (1890), Upham (1903), Peet (1904), and Fairchild (1917, 1919) interpreted these clays as estuarine or marine deposits. Woodworth (1905), Stoller (1911, 1919, 1920, 1922), and Reeds (1927) cited abundant evidence that these were lacustrine deposits. This disagreement sparked a lively debate throughout the early part of this century! The argument was eventually settled in favor of a lacustrine environment.

Eights (1852) observed that the upper, brown clay was covered by yellow, ferrogenous sand. Ries (1890), Peet (1904), and Stoller (1911) also noted that the clays were mantled with yellow, stratified fine sand, and that the sand blanket thickened systematically towards valley-side gravely deltaic deposits. They all interpreted this unit as a shallow-water sediment. Stoller (1911) pointed out that the lake sand was covered with dunes, especially in the Albany-Schenectady area.

Merrill (1890), Ries (1890), Upham (1903), Peet (1904), Woodworth (1905), Fairchild (1917, 1919), and Stoller (1919, 1920, 1922) all used the elevations of the valley-side deltas, wave-cut cliffs, and beaches to define water planes (marine or lacustrine, depending on taste) that tilted up to the north at 2.5 ft/mile (Woodworth, 1905). Woodworth (1905) and Stoller (1919, 1920, 1922) also described multiple lake levels.

The sedimentary facies of Lake Albany can be deduced using the studies discussed above and more recent work by LaFleur (1965a, 1969a), Dineen and Rogers (1979), Dineen and others (1983), and Dineen and Duskin (1987), along with observations of exposures, drill cores, water well logs, and geophysical data. The sedimentary sequence begins with a basal ice-contact facies, which grades up to a deep-water facies, which then grades up into a shallow-water or nearshore facies. This sequence ends with fluvial and/or aeolian deposits.

**Basal Ice-contact Facies:** The base of the Lake Albany sequence generally is in abrupt, unconformable contact with the underlying bedrock, lodgement till, or sand and gravel. Many of the exposures along the Hudson River are of this type. See stops 8, 10, and 13 in Dineen and Rogers (1979) and stop 12 in Dineen and Duskin (1987) for examples of this type of contact. Many water
wells and test borings in the preglacial channels encountered large areas of till or bedrock under the lake clays, and the Basal Ice-contact Facies was missing.

The basal Lake Albany sediments are mostly faulted and deformed cross-bedded sand and gravel with interbedded flowtills. These sand and gravel deposits tend to be fan-shaped, and frequently grade laterally "up-ice" into sinuous gravel ridges or eskers (Dineen and others, 1983). They also grade into valley-side kame delta deposits (LaFleur, 1969a). The ice-contact deposits are subaqueous fans. The ice-contact deposits grade up and laterally into the Deep-water Facies. Exposures that illustrate these relationships include stop 8 in LaFleur (1961b); stop G-5 in Connally and Sirkin (1967); stop 9-5 in LaFleur (1969a); stops 3, 5, 7, and 8 in LaFleur (1979); stops 1, 5, 6, 7, and 16 in Dineen and Rogers (1979); stop 6 in DeSimone and LaFleur (1985); and stops 4, 6, and 9 in Dineen and Duskin (1987). Stops 5, 6, and 7 on this trip also illustrate these features.

Deep-water Facies: The middle portion of the Lake Albany sequence is primarily rhythmically bedded silt and clay. These beds grade up from ripple-laminated silt to rhythmites or "varves" of clay and silt or fine sand. The thickness of each couplet and the proportion of silt-to-clay in them decreases upward. This facies also includes turbidites of ripple-laminated sand or sand and gravel, and lenses or beds of contorted clay, striated boulders, and "rain-out" or dropstone till. The turbidites extend over large areas, based on geophysical interpretation of test borings (Dineen and others, 1983). The "rain-out" till and boulders are probably from icebergs (DeSimone, 1985; DeSimone and LaFleur, 1986) and floating ice shelves (Dineen and others, 1983). The locally contorted clay beds might have resulted from icebergs grounding on the lake bottom, by till impacting the bottom after release from icebergs, and by slumping of unstable bottom deposits. The interpretation of geophysical logs also suggests that the clay units are shingled systematically from south to north, and that the clays have compacted so that they sag towards the axes of the buried bedrock valleys (Dineen and Rogers, 1979; Dineen and others, 1983).

The colors of the clay are clues to the origin of the deposits because the clay is rock flour ground from bedrock surfaces. Distinct reddish clay beds occur in the Hudson Valley adjacent to the Esopus Creek and Cats Kill valleys. Devonian redbeds are exposed in both valleys. The red Cambrian-Ordovician slates in the Batten Kill and Hoosic River drainage basins contributed red clay to the Battenkill-Hudson Gorge next to the Hoosic and Batten Kill deltas. The phyllites in the Hoosic and Batten Kill basins contributed green clays to the easternmost buried channel. White clay occurs near the mouth of the Kayaderosseras Creek, a stream that drained the Adirondack Mountains. The blue or dark gray color of most of the clay deposits is from the dark Ordovician
shales that lie along the Hudson Valley. Oxidation and leaching causes the upper layers of the gray clay to become brown or yellow, and causes carbonate-sulfate concretions to form.

The bedrock topography also controlled the deposition of the clay. Each of the deep, narrow valleys (Fig. 2a) acted as a sediment trap or basin, where thick wedges of silt and clay accumulated. The linear north-south valleys impeded east-west transport and facilitated north to south transport of sediment. The western valleys tend to reflect the contribution of the western tributaries, the central valleys derived most of their sediment from the ice-front, and the eastern valleys received sediment from the tributaries to the east. The basal, ice-contact facies tend to form ridges perpendicular to the ice front that trend northwest-southeast across the buried valleys.

Examples of the Deep-water Facies can be observed at stops 6 and 8 in LaFleur (1961b); stops 12 and 13 in LaFleur (1965b); stop 9-2 in LaFleur (1969a); stops 3, 4, 5, 7, and 8 in LaFleur (1979); stops 1, 2, 12, 13, 14, 15, and 16 in Dineen and Rogers (1979); stops 6 and 7 in DeSimone and LaFleur (1985), stops 4, 6, 11, and 12 in Dineen and Duskin (1987), and stop 7 in Dineen (1987). We will see the Deep-water Facies at Stops 6, 7, and 9.

Shallow-water Facies: The higher wave and current energy in shallow water produced several distinctive types of deposits. The shallow-water deposits began to accumulate as the lake basin filled with sediment and the lake levels fell through time. Thus, they formed in response to the (lake) bottom coming up and the top (water surface) coming down!

Many rock ridges along the Hudson Valley are partially covered with gravelly silt. The gravel is usually quite angular and is derived from local rock. These ridges apparently were above wave-base in the lake, so winnowing of any till mantle or erosion of the rock face could take place.

Wide-spread blankets of yellow brown, fine to medium sand, twenty to forty feet thick, cover the deep-water clay throughout the Hudson Valley (Cadwell and Dineen, 1987). The sand is usually planar- to ripple-laminated and generally has a gradational contact with the underlying clay, except along the edges of the lake plain. It might contain a major contribution of wind-blown, lake-deposited sand and silt, particularly in the Albany area (Dineen and Rogers, 1979). In the Kingston area, the sand is trough-bedded with many cut-and-fill structures (Dineen and Duskin, 1987).

The Shallow-water Facies can be examined at stop 8 in LaFleur (1979); stops 2, 3, 13, and 15 in Dineen and Rogers (1979); and stops 3 and 6 in Dineen and Duskin (1987).
Near-shore Facies: The lake shore and adjacent environments produced several types of deposits and landforms. They include: deltas with topset, foreset, and bottomset beds; fluvial deposits, which include terrace deposits of the later lake stages, and grade into the deltaic topsets; and beaches, including wave-cut platforms. All these sediments were deposited in relatively high-energy environments.

The deltas along the Hudson Valley can be classified as kame deltas, ice-marginal deltas, and ice-free deltas. They all contain trough-cross-bedded sand and gravel topsets, planar-cross-bedded gravelly sand foresets, and ripple-laminated to planar-bedded sand, silt toesets and bottomsets. Glacial Lake Albany delta sizes tend to be greatest in the Albany area and decrease dramatically to the north (Hanson, 1977; Dineen and others, 1983; DeSimone, 1985; DeSimone and LaFleur, 1986; Cadwell and Dineen, 1987). The deltas in the lower lake stages usually increase in size from south to north (Hanson, 1977; Dineen and others, 1983).

Kame deltas comprise outwash derived from the ice front. They have a collapsed ice-contact face on their proximal side. The Schodack, Pollock Road, Rensselaer, Waterford, Hampton, and Newtown Road kame deltas (Dineen and others, 1983) are good examples of this type of delta. They mark the edge of active ice.

Ice-marginal deltas derived their sediment from the highlands adjacent to the glacial lakes. The sediment is predominantly inwash. The collapsed ice contact face is on their distal edge. The Red Hook, Catskill, upper Kinderhook, upper Hoosic, and upper Batten Kill deltas are examples of ice-marginal deltas. Most were deposited next to stagnant ice.

Ice-free deltas lack ice-contact features, except for small kettle-holes. They are most abundant in the lower lake stages. Many of the deltas associated with the smaller tributaries have relatively "starved" topset sequences because many of the later lake stages were short-lived. Glacial ice was almost absent when these deltas were deposited. The Kingston, lower Kinderhook, Schenectady, lower Hoosic, and lower Batten Kill deltas are examples of this type.

Delta exposures include stops 5 and 6 in LaFleur (1961b); stop 13 in LaFleur (1965b); stop 5 in LaFleur (1969a); stops 9-4, 9-5, and 9-7 in LaFleur (1979); stops 1, 2, 5, 9, 10, 11, 12, 13, and 16 in Dineen and Rogers (1979); stops 1 and 4 in DeSimone and LaFleur (1985), and stop 3 in Dineen and Duskin (1987). Stop 7 is in the Pollock Road kame delta.

Fluvial Facies: Fluvial deposits grade into the deltaic topsets. They include terraces that are associated with rivers that fed deltas, terrace deposits inset into pre-existing lake deposits.
during lower lake stages, and major lake outlet or catastrophic flood channel deposits.

Most of the fluvial deposits are trough-cross-bedded sand and gravel. The thickness and the width of the troughs vary widely. These deposits usually underlie terraces that are graded to deltas in adjacent lakes.

Terraces that are incised into older, higher lake deposits are associated with the lower lake stages. They bear an unconformable relationship with the older lake deposits. The terrace sediments are usually poorly-sorted gravelly sand. These sediments are often poorly stratified with horizontal bedding.

Catastrophic floods from Lake Iroquois carved large, deep channels that were graded to the Quaker Springs through Coveville stages of Lake Albany (Stoller, 1911, 1919, 1920; LaFleur, 1965b, 1969a, 1979; Hanson, 1977). These magnificent channels are preserved in the upper Hudson Valley from Schenectady to Schuylerville (Cadwell and Dineen, 1987). Lake Iroquois and its successor, Lake Frontenac, drained through the Covey Hill Spillway into Lake Vermont when the Ontario Lobe retreated north of the Adirondack Mountains (Chapman, 1937; LaFleur, 1965a; Muller and Prest, 1985). Waters escaping from Lake Vermont carved a series of channels near Fort Ann (DeSimone and LaFleur, 1985, 1986) that might include both outlet and catastrophic flood channels.

Large areas of slackwater clay and silt were deposited along the upper edges of catastrophic flood channels from Schenectady to Schuylerville (Dineen, Round Lake and Burnt Hills 7-1/2 minute quads, NYSGS Open File Maps; Hanson, 1977; Dineen and others, 1983). The floods scoured longitudinal grooves in the Fort Ann channels (DeSimone and LaFleur, 1985, 1986), and cut overflow channels across the interfluve between the Ballston and Colonie Channels. Fans of sand and gravel lie at the mouths of the overflow channels in the Clifton Park area (Dineen and others, 1983). The fans are often overlain by slackwater silt.

Lakes Coveville and Fort Ann were broad rivers in the Hudson Valley (LaFleur, 1965a; DeSimone and LaFleur, 1986). Deltas, beaches, spits, and deep-water Coveville deposits have been preserved in the Albany area, however, suggesting that lacustrine conditions continued locally until Lake Coveville time (Dineen and Rogers, 1979; Dineen and others, 1983).

The sediments in the floors and terraces of the outlet and catastrophic flood channels tend to be thin but coarse. Boulder lags are common. Slackwater deposits occur along the channel sides.
Fluvial deposits can be examined at stops 5 and 11 in LaFleur (1961b); stops 4, 8, and 10 in LaFleur (1979); stops 2, 3, and 4 in LaFleur (1983); stop 15 in Dineen and Rogers (1979); and stops 7 and 11 in Dineen and Duskin (1987).

Catastrophic flood or outlet channels can be observed at stops 3, 6, 17 and 18 in LaFleur (1965b); stop 6 in LaFleur (1979); and stop 9 in DeSimone and LaFleur (1985). Stop 1 is in an outlet channel, Stop 8 overlooks the Ballston catastrophic channel, and Stop 10 is in an outlet channel.

Beach Facies: Beach features include planar-bedded gravelly sand, ripple-trough-laminated sand, boulder pavements, and wave-cut scarps. They can be observed at stop 9 in LaFleur (1961b); stops 12 and 14 in LaFleur (1965b); stops 9-4, 9-6, and 9-8 in LaFleur (1969a); stop 1 in LaFleur (1979); and stop 3 in Dineen and Rogers (1979).

Aeolian Facies: Dune Sand overlies many of the sand plains and deltas of the Hudson Valley. It is yellow brown, planar-cross-laminated medium to fine sand. The aeolian sand unconformable overlies the lacustrine deposits. The dunes are parabolic and linear. The cross-bedding and dune morphology suggest that the dune-forming wind came from the northwest (Dineen and Rogers, 1979; Dineen, 1982). Dune sand and dunes can be seen at stops 3 and 4 in Dineen and Rogers (1979).

TERRACES, LAKE PLAINS, AND ISOSTATIC REBOUND

The clay and sand terraces and their associated deltas, beaches, and dunes constitute the primary evidence for several stages of Lake Albany. The water levels of the lake can be deduced from the delta elevations (Fig. 3). Secondary features include beaches, sand and clay terraces, kame terraces, tributary stream terraces, outlet and inlet channels, and catastrophic flood channels. This suite of features allows us to make a first approximation of the sequence of lake stages and of their relative ages and post-glacial rebound (Table 1).

The "water planes" defined by the terraces, beaches, and deltas of the highest stage of Lake Albany tilt up to the north at 2.6 ft/mile to 2.7 ft/mile (0.49 to 0.51 m/km; DeSimone and LaFleur, 1986). Earlier estimates of tilt ranged from 2.25 ft/mile (0.43 m/km; Chadwick, 1928) to 2.5 ft/mile (0.47 m/km; Woodworth, 1905; Fairchild, 1917; LaFleur, 1965a). The water planes of the lower stages are tilted to a lesser degree—those of Lakes Quaker Springs and Coveville are tilted at 1.5 ft/mile (0.28 m/km; LaFleur, 1965a) to 1.6 ft/mile (0.3 m/km; DeSimone and LaFleur, 1986). The Fort Ann stage tilts from 1.05 ft/mile to 1.0 ft/mile (0.2 to 0.19 m/km; DeSimone and LaFleur, 1986). Thus, the water planes of the various lake stages converge to the south (Fig. 3).
GLACIAL LAKES
1. LAKE HARLEM
2. LAKE HACKENSACK
3. LAKE ALBANY
4. LAKE QUAKER SPRINGS
5. LAKE COVEVILLE
6. LAKE FORT ANN

DELTAS
• Ice-free
• Ice-marginal
• Kame

Figure 3. LAKE TERRACES
Woodworth (1905), Fairchild (1917), LaFleur (1965a), and Connally and Sirkin (1986) noted "hinge lines" on the lake planes at the Hudson Highlands and Glens Falls, where the terrace levels seem to undergo sharp changes in grade. The terraces are less steep or flatten out south of the Highlands (Fig. 3) and become significantly steeper north of Glens Falls (DeSimone and LaFleur, 1986).

The multiplicity of deltas and beaches define many lake stages in the Hudson Valley (Fig. 3). The earliest stage is the highest, and the lakes become younger with decreasing elevation. The highest stage is defined by both ice-marginal and kame deltas, and is clearly a proglacial lake (Woodworth, 1905; Fairchild, 1917; Chadwick, 1928; LaFleur, 1965a and b; Connally and Sirkin, 1986). A lower, somewhat ice-free Lake Albany stage has been mapped in the Albany area (LaFleur, 1965a; Dineen and others, 1983). It is characterized by fewer kame deltas (Fig. 3). This level was considered the "Stable Lake Albany" stage by Connally and Sirkin (1986). Ice-marginal deltas persist until the Lake Coveville stage (Fig. 3). Lake Coveville has not been split into substages. DeSimone And LaFleur (1986) were able to divide the Fort Ann Stage into three levels based on three overflow channels for the Lake Vermont-Fort Ann stage in the vicinity of Glens Falls, NY. They confirmed the fluvial nature of the Fort Ann stage in the Hudson Valley. The scatter of points representing deltas and beaches suggest that several intermediate "stable" substages probably exist for Lakes Quaker Springs and Coveville (Fig. 3).

The uppermost, proglacial Lake Albany stage was thought to be a product of a "peripheral bulge" that formed in front of the retreating ice sheet (Peet, 1904). The peripheral bulge was caused by a very rapid isostatic uplift of the earth's crust that formed a few tens of miles in front of the ice tongue. Fairchild (1917) considered the bulge to be a wave movement that was 100 miles wide and 100 feet high. Chadwick (1928) presented a theoretical argument for the presence and character of a peripheral bulge (see figs. 11 through 16 in Chadwick, 1928). Connally and Sirkin (1986) resurrected the peripheral bulge idea and described the upper Lake Albany stage as a very short-lived, very small lake that formed along or on the retreating ice margin. Unfortunately, no one has been able to predict unique geomorphic or stratigraphic features that would prove the existence of a bulge.

Lake Albany levels have been confidently traced from Glens Falls, NY, to Newburgh, NY (Woodworth, 1905; LaFleur, 1965a). Lake terraces have been mapped south of the Hudson Highlands (Reeds, 1927), although terraces within the Hudson Highlands have been shown to be underlain by till (Peet, 1904; Woodworth, 1905). The lakes south of the Highlands were named Lakes Hackensack and
Hudson by Reeds (1927). Woodworth (1905) thought that an ice margin at the north edge of the Hudson Highlands was the northern limit of Lake Hudson. Connally and Sirkin (1986) correlated Lake Hudson with their "Stable Lake Albany" stage.

THE OUTLET OF LAKE ALBANY

The northern end of Lake Albany has been fairly well mapped, whereas its southern end has not. The dam or spillway that controlled the lake levels has never been identified in the field. As discussed above, Peet (1904) and Fairchild (1917) thought that the lake was dammed by a peripheral bulge. Woodworth (1905) thought that Lake Albany was dammed by a recessional moraine in the Hudson Highlands and considered the till terraces in the Hudson Highlands gorge to be remnants of the recessional moraine. Reeds (1927) and Connally and Sirkin (1986) suggested that the dam was an extension of the Terminal Moraine between Staten Island and Long Island. The lack of evidence for a dam was attributed to the postglacial erosion of the earthen Terminal Moraine or recessional moraine (Woodworth, 1905; Reeds, 1927; Connally and Sirkin, 1986). Severe, late glacial fluvial erosion occurred in the lower Hudson Valley during and after the Quaker Springs stage of Lake Albany (Merrill, 1890; Peet, 1904; Woodworth, 1905; Newman and others, 1969; Lovegreen, 1974; Dineen and Duskin, 1987). Many of the clay deposits between Poughkeepsie and Long Island were deeply scoured and overlain by fluvial sand and gravel.

Several lakes occupied the lower Hudson Valley. They included Lake Hackensack in the Triassic Lowlands of New Jersey and New York, Lake Hudson in the Hudson Gorge, and Lake Flushing in Long Island Sound (Figs. 1c and 2b; Reeds, 1927). Only Lake Hackensack shorelines are well preserved, although several kame deltas or ice-marginal deltas lie at the Albany level. The Lake Albany features include a delta at 30 ft (9.1 m) near Maurer, NJ (Reeds, 1927), a kame delta at Croton, NY, at an elevation of 100 ft (30.5 m; Markl, 1971), and lake terraces between 100 and 120 ft (30.5 and 36.6 m) near Nyack, NY (Peet, 1904; Woodworth, 1905). All three lakes were dammed by the Terminal Moraine. Their water planes dip up to the north at 2.25 ft/mile (0.43 m/km; Fig. 3; Reeds, 1927). The Lake Hackensack water plane projects 40 ft (12.2 m) above the water plane of Lake Albany in the Sparkill Gap area (compare Figs. 3 and 5a). Thus, Lakes Albany and Hackensack are not the same lake.

The projection of the Lake Albany water plane lies 50 ft (15.2 m) above the floor of the notch in Sparkill Gap near Piermont NY (Figs. 3 and 5; Averill and others, 1980; Dineen, 1986b). Sparkill Gap is a broad windgap that is 12,200 ft (3.7 km) wide and 420 ft (128 m) below the elevation of the adjacent Palisades Escarpment (Figs. 3 and 5). The gap has two parts, a broad strath terrace at an elevation of 190 ft (58 m), and a narrow inner
gorge or notch at the gap's north end (Fig. 5a). The gorge is V-shaped, 1,600 ft (490 m) wide, 160 ft (48.8 m) deep, and is floored with silty fine sand. Sparkill Creek presently drains east through the gorge. The divide between Sparkill Creek and the Hackensack Basin lies in a low-relief gravel-floored channel southwest of Sparkill Village.

Seismic refraction lines and test boring data show a bedrock sill or lip at 30 ft (9.1 m) below sea level in the gorge. The gorge hangs on the buried Hudson River gorge (Newman and others, 1969; Averill and others, 1980). The deep, buried Piermont channel extends southwest from the Sparkill gorge into the Hackensack Lowlands, where it meets the preglacial Oradell channel (Perlmutter, 1959; Lovegreen, 1974).

The Oradell channel and the Piermont channel are filled with glacial sediments, including red silt and clay of Glacial Lake Hackensack (Lovegreen, 1974; Averill and others, 1980). The Hackensack silt and clay have been truncated and subaerially oxidized between Oradell, NJ, and Elizabeth, NJ (Lovegreen, 1974). These silts and clays are overlain by trough-cross-bedded sand and gravel. The sand and gravel underlies a 30-foot (9.1 m) channel and a 50-foot (15.2 m) terrace in the vicinity of Sparkill (Stops 1a and b). The terrace slopes down to 30 ft (9.1 m) ASL (Above Sea Level) at Harrington, NJ, where the channel surface lies at an elevation of 25 ft (7.6 m) ASL (Fig. 5b). The sand and gravel deposit can be traced south into the Hackensack Lowlands, where it overlies the dissected Lake Hackensack deposits (figs. 15 through 18 in Lovegreen, 1974; fig. 15 in Averill and others, 1980). The sand and gravel grades into reddish brown, massive to planar-bedded, silty sand to sandy silt at the Sparkill Gap gorge.

The 190-foot (58 m) strath terrace at Sparkill is the Pliocene channel of the Hudson River (Lovegreen, 1974). The inner gorge or notch at Sparkill is the spillway of Lake Albany (Dineen, 1986b). Water debouching through the spillway from Lake Albany scoured the emergent Lake Hackensack deposits and deposited the trough-cross-bedded sand and gravel.

ICE MARGINS AND THE QUESTION OF READVANCES

Recent mapping has revealed a number of ice margin positions in the Hudson Lowlands (Fig. 4; LaFleur, 1965a and b; Connally and Sirkin, 1967, 1970, 1971, 1986; Hanson, 1977; Dineen and others, 1983; DeSimone, 1985; DeSimone and LaFleur, 1985, 1986; Dineen, 1986a; Dineen and Duskin, 1987; Cadwell and Dineen, 1987). These ice margins are marked by kame or till moraines, kame deltas, heads-of-outwash, and the chronologies of tributary lake basins (see the discussions for Stops 11, 12, and 13).
Figure 4. ICE MARGINS AND CARBON LOCALITIES
(See next page for key)
FIGURE 4: KEY

A: Terminal-Ronkonkoma Moraines
B: Harbor Hill Moraine
C: Sparkill-Tappan Moraines
D: Ogdensburg-Culvers Gap Moraines
E: Augusta Moraine
F: Pellets Island Moraine
G: Shenandoah Moraine
H: New Hampton Moraine
I: Poughkeepsie Moraine
J: Wallkill Moraine
K: Hyde Park Moraine
L: Wagon Wheel Gap Moraine
M: Whitfield Moraine
N: Red Hook Moraine
O: Prattsville-Stone Ridge-Cottekill-Ulster Park Moraines
P: Bell Pond Moraine
Q: Middleburg Moraine
R: Jackson Summit-Yost-Vly Moraines
S: Broadalbin-Perth-Meadowdale-Schodack Moraines
T: Hampton-Wynantskill-Poestenkill Moraines
U: McKownville-Loudenville-Rensselaer-Sycaway Moraines
V: Niskayuna-Tomhannock-Hoosick Falls Moraines
W: Willow Glen-Hale Mountain Moraines
X: Randalls Corners-Arlington-Woodland Lake Moraines
Y: Hidden Valley-Black Valley-Patten Mills Moraines
Z: Welch Hollow Moraine
FIGURE 5a- SPARKILL GAP CROSS SECTION

FIGURE 5b- SPARKILL GAP
While we enjoy some unanimity on the locations of the ice margins, we disagree on their meaning. Some workers have suggested that the margins were deposited by advancing ice (Connally and Sirkin, 1970, 1971, 1986; Dineen and others, 1983; Dineen, 1986a) while others have insisted that they are the product of systematically retreating ice (Duskin, 1985; DeSimone and LaFleur, 1985, 1986). The evidence for large-scale readvances (on the order of tens of miles) should include widespread till sheets over folded or thrust lacustrine or fluvial deposits, rejuvenation of outwash and tributary lake systems, and multiple tills. Local exposures of contorted clays or of till-over-something abound (Connally and Sirkin, 1967, 1970; Hanson, 1977; Dineen and Rogers, 1979; Dineen and others, 1983; Dineen and Duskin, 1987), but continuous till sheets or other mappable features have not been found. The available surface and subsurface data is ambiguous (DeSimone and LaFleur, 1985, 1986; Duskin, 1985; Dineen and Duskin, 1987), but suggests that the ice margin oscillated. Recent detailed mapping in the Hudson Valley (DeSimone, 1985; Duskin, 1985) has contradicted the existence of the long-established Rosendale (Connally and Sirkin, 1970), Delmar (Dineen, 1986a), and Luzerne Readvances (Connally and Sirkin, 1971).

Two areas do seem to have experienced (minor?) readvances, however. Till occurs over compacted "varved" clay in the test borings for the Newburgh-Beacon Bridge and highly compacted gravel underlies uncompacted clay at the Mid-Hudson Poughkeepsie Bridge (Dineen and Duskin, 1987). Till-over-clay or -outwash has been encountered in water wells and test borings in the Casper Creek (Stop 3; Hanson, various unpublished reports for Dunn Geoscience Corp.) and Sprout Creek Valleys (Moore, 1982). Till-over-clay and -outwash has also been observed in the Wall Kill Valley (Russell Waines, personal communication). These deposits might have been emplaced by a readvance from Poughkeepsie to the north flank of the Hudson Highlands, or could be pre-Woodfordian deposits that were overridden by the Woodfordian glacier.

The second area of till-over-something is in the Helderberg Plateau, where till overlies gravel or clay throughout the Cats Kill Valley (Dineen, 1986a). The till was probably deposited by the Middleburg Readvance (LaFleur, 1969b). These deposits have not been correlated with events in the Hudson Lowlands, however, and might be pre-Woodfordian sediments.

Interpretation of the available data suggests that a conservative approach to the question of readvances is best. At present, the evidence suggests that the ice margin oscillated somewhat, but the magnitude of the oscillations is unknown. Perhaps the first step would be to develop a model of the expected deposits and features that would be associated with a glacial readvance or oscillating glacial margin in a proglacial lake, and then testing the models with detailed field work.
FOSSILS, RHYTHMITES, AND RADIOCARBON DATES

The Lake Albany sediments are relatively fossil-poor. Ries (1890) mentioned finding a fragment of a diatom in the Hudson Valley clay. Peet (1904) described sponge spicules, fresh water diatoms, and insect tracks in clay at Croton and leaves of Vaccinias oxyccoccus in clay at Albany. Hartnagel and Bishop (1922) noted several tantalizing fossil sites, including the Cohoes mastodon, found in a pothole 100 feet above the Mohawk River at Cohoes. They also mention a find at lock 6 of the Champlain Canal, near Fort Ann, where cones of Canadensis and spagnum moss were unearthed. Woodworth (1905) cites Asa Fitch's notes that an ash, beechnut, and butternut-rich bed was encountered beneath fluvial(?) sand in the Wood Creek Valley (between the Champlain and Hudson Lowlands) during the construction of the Champlain Canal. Fitch was very impressed by the fact that the trees were all tipped over, with their tops pointing south, opposite the present-day flow of Wood Creek. A Rangifer arcticus (caribou) antler was found in a gravel pit in the Schenectady Delta (Fisher and Ostrom, 1952).

Several fossils have been found recently. In 1977, Tom Engel (NYS Department of Environmental Conservation) found fossiliferous rhythmites in the excavation for the sewer treatment plant at Glenville Rest Home near Schenectady, NY. The fossils were mollusks that were in clay infilling the Ballston channel to an elevation of 240 ft (73.2 m). The attached report by Eileen Jokinen describes the fossils (Appendix 1). Gail Ashley was able to identify Lebensporen (insect larva tracks) on silt laminae surfaces at Stop 5 (personal communication, 1986).

The Deep-water Facies is a series of silt-clay couplets 0.4 to 6 inches (1 to 15 cm) thick (Dineen and others, 1983). The couplet's base is usually on an eroded surface, and consists of planar to ripple-laminated silt or sandy silt that grades up to clay (Dineen, 1977). The percentage of silt systematically decreases upward (Dineen, 1977). The silt-rich couplets are thicker close to deltas or the ice front. Dineen (1977) measured a section with 600 rhythmites near Albany. The average rhythmite thickness was 0.5 to 2 inches (1.3 to 5 cm) in this section. Extrapolating this average thickness, the section encountered in the deepest part of the Colonie Channel (Bore Hole 5 in Dineen and others, 1983) contained 5,800 to 1,450 couplets. Engineers and the early mappers identified the couplets as "varves." LaFleur (1965a) did not find compelling evidence that they were annual accumulations of sediment, and suggested that the more-neutral term "rhythmites" be used. If the rhythmites are true varves, Lake Albany lasted 1,450 to 5,800 years in the Albany area.
Table 2 summarizes fifteen radiocarbon dates from the Hudson Valley. Although there is some scatter in the dates, and the extrapolation from the age of the base-of-organic-rich sediment to the age of base-of-the-bog is fraught with peril (Dineen, 1986a), the systematically younger dates from south to north suggests that the Lake Albany stage lasted from 17,000+ to 13,200 yBP. The lake stages from Quaker Springs through Fort Ann lasted from 13,200 yBP to the inception of the Champlain Sea episode. Date Number 1 on Table 2 suggests that the Champlain Sea episode began prior to 10,300 yBP. Clark and Karrow (1984) favored a date of 12,100 yBP for Lake Iroquois, and 11,620 yBP, based on a pool of seven radiocarbon dates on marine shells in the St. Lawrence Valley, for the Champlain Sea. Muller and Prest (1985) favored a date between 11,800 and 10,630 yBP (11,200 yBP?) for the Champlain Sea.

REGIONAL STRATIGRAPHY

Figures 6, 7, and 8 depict nine phases in the history of Lake Albany. Many other phases can be considered using the ice margins determined by Dineen (1986), Connally and Sirkin (1973, 1986), and DeSimone and LaFleur (1986), and the lake stages depicted in Figure 3 and DeSimone and LaFleur (1986). Table 1 presents tentative correlations with the Hackensack, Ontario, and St. Lawrence Lowlands.

FIGURE 6a: Ogdensburg–Culvers Gap–Sparkill Moraines, Glacial Lakes Wall Kill, Hackensack, and Hudson–Flushing: The earliest phase depicted is the retreat from the Terminal Moraine and the development of Lake Hackensack, with concurrent formation of the 600-foot (180 m) stage of Glacial Lake Wall Kill, the Ogdensburg–Culvers Gap Moraine, Lakes Hudson and Flushing, and the Sparkill Moraine (Figs. 2b and 4). The Connecticut Lobe blocked Long Island Sound as the Hudson Lobe retreated. This happened approximately 19,000 years ago (Sirkin, 1977, 1986).

FIGURE 6b: Cuddebackville–New Hampton–Shenandoah Moraines, Glacial Lakes Wall Kill and Hudson–Flushing: As the Hudson Lobe retreated to the north slope of the Hudson Highlands, it deposited the Cuddebackville, New Hampton, and Shenandoah moraines (Fig. 4). Lakes Wall Kill and Hudson–Flushing were trapped between the ice front and the Terminal Moraine (Fig. 2b). Lake Hackensack drained due to rapid isostatic uplift (perhaps a peripheral bulge), and Lake Hudson began to overflow through Sparkill Gap. The overflow eroded the emergent Hackensack clays. The readvance (?) features between Poughkeepsie and the Highlands might have been formed at this time.

TABLE 2  RADIOCARBON DATES

<table>
<thead>
<tr>
<th>NUMBER ON FIG. 4</th>
<th>LOCALITY</th>
<th>RADIOCARBON DATE (YrB.P.)</th>
<th>BOG BOTTOM AGE*</th>
<th>LARP NUMBER</th>
<th>REFERENCE &amp; SIGNIFICANCE</th>
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<tbody>
<tr>
<td>4</td>
<td>New Hampton Bog No. 1</td>
<td>12,510±370</td>
<td>---</td>
<td>I-4137</td>
<td>Steadman and Funk, this manuscript. Dates caribou bone in Wallkill Valley.</td>
</tr>
<tr>
<td>3</td>
<td>Dutchess Quarry Cave</td>
<td>12,850±250</td>
<td>---</td>
<td>L-1157a</td>
<td>Connally and Sirkin, 1970. Dates the base of organics. Dates the retreat from the Wallkill Moraine.</td>
</tr>
<tr>
<td>2</td>
<td>Oradell Reservoir</td>
<td>12,150±210</td>
<td>---</td>
<td>QC-297</td>
<td>Averill and others, 1986. Dates the organic sediment overlying a mastodon. Dates the organic accumulation in Sparkill-Oradell channel.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>12,870±200</td>
<td></td>
<td>QC-296</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>West 50th St.</td>
<td>10,280±270</td>
<td>---</td>
<td>J-5200</td>
<td>Weiss, 1974. Dates the base of organics. Dates the flooding of Hudson Estuary.</td>
</tr>
</tbody>
</table>

*Assuming an accumulation rate of 0.036 cm/yr for inorganic sediment (Connally and Sirkin, 1986).
### TABLE 2 RADIOSCARBON DATES

<table>
<thead>
<tr>
<th>NUMBER ON FIG. 4</th>
<th>LOCALITY</th>
<th>RADIOCARBON DATE (YrBP)</th>
<th>BOG BOTTOM AGE*</th>
<th>LAB NUMBER</th>
<th>REFERENCE &amp; SIGNIFICANCE</th>
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<tr>
<td>8</td>
<td>Meadowdale Bog</td>
<td>10,485±324 16,650±660</td>
<td>14,790</td>
<td>GX-8487</td>
<td>Dineen, 1986. Younger date is from base of organics, older date is from organic-poor base of core. Dates retreat from the Meadowdale Margin.</td>
</tr>
<tr>
<td>7</td>
<td>Great Bear Swamp</td>
<td>11,590±265 19,875±980</td>
<td>13,950</td>
<td>OC-149</td>
<td>Dineen, 1986. Older date is from organic-poor base of core. Dates retreat from the Helderberg Plateau.</td>
</tr>
<tr>
<td>5</td>
<td>Poughkeepsie IBM</td>
<td>10,310±85 10,505±100 13,210±550 13,260±125</td>
<td>---</td>
<td>SI-6016 SI-6016a GX-9435 SI-6030</td>
<td>Connally (personal communication). Dates material in upper section of folded clay, under trough-cross-bedded sand and gravel. Dates fall to Lake Quaker Springs or Coveville.</td>
</tr>
</tbody>
</table>

*Assuming an accumulation rate of 0.036 cm/yr for inorganic sediment (Connally and Sirkin, 1986).*
Figure 6a.
LAKE HACKENSACK
SPARKILL-OGDENSBURG-CULVERS
GAP MORAINES

Figure 6b.
LAKE HUDSON
NEW HAMPTON MORAINES

Figure 6c.
LAKE ALBANY
WAGON WHEEL GAP MORAINES
ago (Fig. 4). The moraines of this position are well developed, implying a relatively lengthy stillstand of the ice. Lake Wallkill fell to its 400-foot (120 m) level and spilled eastward through the Moodna Creek Valley, while Lake Hudson-Albany expanded into the mid-Hudson Valley.

FIGURE 7a: Middleburg-Whitfield-Cottekill-Ulster Park-Red Hook Moraines, Glacial Lakes Grand Gorge-Schoharie, Warwarsing, Tilson, Albany, and Attlebury: The Hudson Lobe retreated to the northern rim of the Helderberg Plateau before advancing to the Middleburg-Whitfield-Cottekill-Ulster Park-Red Hook margin (Fig. 4). Lakes Schoharie, Warwarsing, Tilson, and Attlebury formed along the margin (Fig. 2). All the lakes (except the earliest stage of Lake Schoharie) spilled into Lake Albany. The ice retreated from this margin about 16,000 years ago (Table 2). This event probably correlates with the Valley Heads readvance into the western Mohawk Valley (Muller and others, 1986). Note that DeSimone does not agree with the Middleburg readvance hypothesis.

FIGURE 7b: Spruce Mountain-Meadowdale-Hampton-Wynantskill-Poestenkill-Pownal Moraines, Glacial Lakes Sacandaga, Schoharie-Amsterdam, Albany, Tomhannock, and Bascom: The Hudson Lobe had retreated to the Spruce Mountain-Meadowdale-Hampton-Wynantskill-Poestenkill-Pownal ice margin by 15,000 years ago (Table 2, Fig. 4). Lake Sacandaga formed in the Sacandaga Basin (Fig. 2b). Glacial Lake Schoharie drained into the Cats Kill Valley through the Franklinton notch (Fig. 1c). As the Mohawk Lobe retreated from the Mohawk Valley and exposed the Duanesburg-Bozen Kill outlets, Lake Schoharie fell to the Lake Amsterdam level (Figs. 1c and 2b). Lake Amsterdam filled the lower Mohawk Valley and eventually spilled through the West Hill Gorge, while Lakes Tomhannock and Bascom expanded in the Taconic Mountains section of the New England Upland.

FIGURE 7c: Hidden Valley-Glen Lake-Carter-Manchester Moraines, Glacial Lakes Warrensburg and Albany: As Lake Albany expanded to its maximum size, the glacier retreated into the Champlain Valley. Lake Warrensburg flooded the Upper Hudson River Valley (Fig. 2b). The Mohawk Valley carried relatively steady meltwater flow from western New York (Fig. 1c).

FIGURE 8a: Glacial Lake Vermont-Lake Quaker Springs stage: The Hudson Valley became ice-free by 13,100 years ago (Table 2). The Hudson Valley lake level fell to the Quaker Springs stage at the same time that catastrophic floods from the Appalachian Plateau, Ontario Basin, and the Adirondack Highlands began to flush over the Rome Outlet and through the Mohawk Valley (Fig. 1c). The floods carved channels between Schenectady and Saratoga Lake, and might have caused the transition from the Albany to the Quaker Springs stages.
FIGURE 8b: Glacial Lake Vermont-Coveville stage: Lake Iroquois began to fill the eastern Ontario Basin, and Lake Vermont expanded in the Champlain Valley as the Hudson Valley lake fell to the Coveville level (Fig. 1c). The Coveville stage was fluvial-lacustrine throughout the Hudson Valley. Catastrophic floods still came through the Mohawk Valley from the Ontario Basin and the Adirondacks. These floods might have triggered the drop in lake levels from the Quaker Springs to the Coveville stage, and the drop from the Coveville to the Fort Ann stage.

FIGURE 8c: Glacial Lake Vermont-Fort Ann stage: The Sparkill Gap was abandoned as Lake Coveville fell to Lake Fort Ann. Catastrophic floods severely eroded the lower Hudson Valley following the breaching of the Terminal Moraine. Riverine conditions dominated the Hudson Valley from Fort Ann through the Terminal Moraine. The Rome Outlet was abandoned as the retreating ice uncovered the Covey Hill Spillway and Lake Iroquois fell to Lake Frontenac. Lake Iroquois and post-Iroquois flood events might have caused the successively lower Fort Ann stages.

The Hudson River began to assume its present character between 11,000 and 11,500 years ago, when the ice front pulled back enough to allow the Champlain Sea to invade the St. Lawrence Lowlands and the supply of meltwater to the Hudson Valley was cut off.

SUMMARY AND CONCLUSIONS

Lake Albany lasted from ±18,000 to 13,200 yBP and extended from the Terminal Moraine, between Staten and Long Islands, to Glens Falls, NY. The later Lake Albany stages ended 11,250 yBP. Lakes Hudson and Albany were continuous and outlasted Lake Hackensack. Lake Hudson-Albany was dammed by the Terminal Moraine, but overflowed or spilled through Sparkill Gap into the Hackensack Lowlands. The overflow from the lake eroded the emergent Lake Hackensack clays, and deposited a ribbon of fluvial sand and gravel from Sparkill, NY, to Newark, NJ.

The water planes of the stages of Lake Albany can be deduced by correlating sand and clay plains, deltas, and beaches. Several lake stages can be documented using these features, including Lake Albany I and II (ice-contact and stable), Lake Quaker Springs, Lake Coveville, and Lake Fort Ann I, II, and III. The water planes of the various stages tilt down to the south because of glacial isostatic rebound. The tilts decrease from oldest (Lake Albany I) to youngest (Lake Fort Ann III). The water plane of Lake Albany tilts 2.6 ft/mile (0.49 m/km), while that of Fort Ann III tilts 1.1 ft/mile (0.21 m/km). Lakes Quaker Springs, Coveville, and Fort Ann I, II, and III project into the Champlain Valley, where they were stages of Lake Vermont. The Coveville and Fort Ann stages were riverine in the Hudson Valley.
The facies of the Lake Albany deposits change systematically from ice-contact sand and gravel at the base to rhythmically laminated silt and clay in the middle, distal beds, to silty sand in shallow water sediments, to gravelly sand in near-shore sediments. The sediment sequence was strongly influenced at the base by the retreating Hudson Glacial Lobe, and at the top by the shoaling effects of falling water levels. Lake Albany deposits are sparsely fossiliferous. Mollusks and trace fossils have recently been found. The fossils that are present suggest that the water level was stable for decades, and that vegetation and fish were present.

Moraine segments, kame deltas, and heads-of-outwash document many recessional ice margins in Lake Albany. Possible readvances occurred in the Poughkeepsie-Newburgh area (un-named) and on the Helderberg Plateau (Middleburg Readvance). Retreating ice margins include the Terminal Moraine, the Wagon Wheel Gap-Wallkill, the Meadowdale, and the Lake Juarezne-Carter-Manchester margins. The Terminal Moraine became the dam for the lake. The Wagon Wheel Gap margin is marked by extensive moraines, and was probably deposited during a major stillstand. The Meadowdale margin coincided with the formation of Lake Sacandaga in the southern Adirondack Uplands, Lake Amsterdam in the Mohawk Valley, and Lake Bascom in the Taconic Mountains. The Carter margin was the last ice front in Lake Albany, coincided with the development of Lake Warrensburg in the eastern Adirondack Mountains, and marked the last mapped recessional position in the Hudson Valley.

The Quaker Springs and Coveville lake stages received several (two or three) catastrophic floods from the Mohawk Valley. Later, during the Fort Ann stages, two or three more catastrophic floods debouched from the Champlain Valley into the Hudson Lowland. The catastrophic flood events seem to have coincided with the successive drops in lake levels. The Terminal Moraine dam probably collapsed during a catastrophic flood during the Fort Ann stage.

If the Middleburg readvance is Woodfordian in age, it probably correlates with the Valley Heads readvance in Central New York. Lakes Quaker Springs and Coveville were contemporaneous with Lake Iroquois, and Lake Fort Ann existed during Lake Frontenac time. The Hudson River attained its present drainage area when the Hudson-Champlain Lobe retreated from the Covey Hill spillway and the Champlain Sea flooded the St. Lawrence and Champlain Lowlands.

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APPENDIX

Molluscan fauna of Glacial Lake Albany

by Eileen H. Jokinen, Visiting Associate Professor, Biological Sciences Group, The University of Connecticut, Storrs, CT 06226

The site of the Glacial Lake Albany mollusks is in the excavation for the Glenville Homes Sewer Treatment Plant. The excavation was approximately 15 ft (4.6 m) deep. Its top was at an elevation of 240 ft (73 m; based on the Schenectady 7-1/2 minute quadrangle map). The clays were truncated by catastrophic floods coursing up the Alplaus-Ballston channel. The rhythmically laminated clays and silts were probably distal bottomset beds of the 340 ft (104 m) Schenectady delta. The bivalves were found in "life" position by Mr. Tom Engel (R. J. Dineen, personal communication). The fossils are in the Collections of the New York State Museum.

Class Gastropoda

Subclass Prosobranchia

Family Valvatidae
Valvata tricarinata (Say, 1817)
Valvata sincera (Say, 1824)

Family Hydrobiidae
Probythinella lacustris (Baker, 1928)

Amnicola sp.

Subclass Pulmonata

Family Lymnaeidae
Fossaria galbana (Say, 1825)

Family Planorbidae
Helisoma anceps (Menke, 1830)
Gyraulis cf. altissimus (Baker, 1919)

Class Bivalvia

Sphaeriacea -shell fragments
Unioniacea -shell fragments

RECENT DISTRIBUTION AND ECOLOGY

Valvata tricarinata: Distribution: Northern North America from New Brunswick to Virginia, west to Iowa and Nebraska, northwest to the Northwest Territories south of the tree line (Clarke, 1973, 1981)

Ecology: In Connecticut, V. tricarinata is found only in permanent hard water ponds and lakes (Jokinen, 1983). In other parts of its range, it may be present in slow rivers, streams,

Valvata sincera: Distribution: Clarke (1981) lists three species of Recent sincara: V. sincera ontariensis (Baker, 1931); V. sincera sincera (say, 1924); and V. sincera helicoidea (Dall, 1905). Each has a different distribution:

- **ontariensis**: only from Lake Superior (Shakespeare Island Lake) and the region north of Lake Superior drained by the headwaters of the Attawapiskat, Albany, and Seven river systems. The Lake Albany fossils are not this subspecies but could be either one of the following:

- **sincera**: Newfoundland to British Columbia and the Yukon Territory, and Maine to Minnesota.

- **helicoidea**: more northern than sincera: from Labrador to British Columbia, and Alaska north of a line running from Ungava Bay to southern Victoria Island in the Arctic archipelago. Principally an Arctic and Subarctic species (Clarke, 1981).

Ecology: V. sincera sincera prefers cold, larger lakes in the southern parts of its range (Baker, 1928) but may also occur in rivers, permanent ponds, and muskeg pools (Jokinen, 1983; Clarke, 1981).

- **V. sincera helicoidea** occurs in lakes, ponds, slow-moving rivers and streams, and muskeg pools, usually among aquatic vegetation (Clarke, 1981).

Probythinella lacustris: Distribution: New York west to the Dakotas, Iowa, Nebraska; south to Kentucky, Arkansas, Missouri, Alabama; in Canada from Quebec to the Northwest Territories and Alberta (Baker, 1928; Hibbard and Taylor, 1960; Clarke, 1981).

Ecology: P. lacustris lives in permanent lakes, ponds, and rivers on vegetation and/or on muddy, sandy, or marly bottoms. In the southern parts of its range, the snails live in deeper water (Berry, 1943; Clarke, 1981).

Fossaria galbana (called F. decampi in Clarke, 1981): Distribution: Great Lakes-St. Lawrence drainage north to the Hudson Bay lowlands and the Mackensie River, and west to British Columbia. The southern limits have not been determined (Clarke, 1981). This species has not been found in Connecticut (Jokinen, 1983).

Ecology: A cold water species occurring only in large lakes in the southern part of its range, and in both lakes and rivers in the northern. It lives on submerged vegetation (Clarke, 1981).

Helisoma anceps: Distribution: North America from Georgia north to northern Canada south of the tree line, from the east coast to

Ecology: This is one of the most common species in southern New England today. Populations are most common in ponds greater than 10 hectares but also live in smaller lentic habitats as well as rivers and streams. Commonly found on decaying, submerged, terrestrial leaf litter. The species has a wide tolerance for water chemistry with Connecticut pH ranging from 5.8 to 8.9 (Jokinen, 1983). In Canada, this species lives in lakes, ponds, rivers, and streams among vegetation and on various substrates (Clarke, 1981).

Gyraulis cf. altissimus: Distribution: Listed by Baker (1928) as only known as a fossil. Found in Pleistocene deposits in Ohio, Indiana, Illinois, Michigan, Wisconsin, and Connecticut (Baker, 1928; Cooper, 1930).

CONCLUSIONS

The presence of the valvatids and Probythinella lacustris indicates that the lake maintained a fairly stable water level for at least decades or centuries. Macrophytic vegetation should also have been present for protective cover and as a substrate for diatoms and other algae upon which the snails feed. The abundance of clams in the "life position" implies both a viable in situ habitat and the presence of fish hosts for the unioniacean clam larvae.

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FIELD TRIP STOPS

The field trip stops will be in the Hudson Valley and western Taconic Mountains. The southernmost stop will be at Sparkill Gap, in the outlet for Lake Albany, and the easternmost stops will examine the deposits of Glacial Lake Bascom. The field stops are located on Figure 9.

STOP 1- SPARKILL GAP: Sparkill Gap is a windgap in the Palisades Ridge at Piermont, in the Nyack 7-1/2 minute quadrangle of Rockland County, NY. The glacial and bedrock geology of Rockland County was described in Perlmutter (1959). The bedrock topography is strongly controlled by the relative hardness of the Triassic bedrock. Strike ridges have developed on southwest-dipping sandstones and conglomerates. Shale beds underlie the adjoining strike valleys. The Palisades Escarpment is the most prominent ridge in the area (Fig. 5b). The Palisades are underlain by a sill of highly resistant, westward dipping diabase.

The broad Sparkill Gap is 12,200 ft (3.7 km) wide and lies 420 ft (130 m) below the general elevation of the Palisades Ridge (Figs. 5a and b). The Sparkill Gap consists of a wide strath terrace at an elevation of 190 ft (58 m), and a northern, V-shaped inner gorge with a floor at 15 to 20 ft. (4.6 to 6.1 m). The gorge is 1,600 ft (490 m) wide at its top, 400 ft (122 m) wide at its bottom, and 160 ft (49 m) deep. It is filled with silty sand to an elevation of 30 ft (9.1 m) below sea level. The inner gorge overlies a northeast-southwest trending fault in the diabase ridge (Perlmutter, 1959).

A series of kames, kame terraces, and kame moraines lie west of the Sparkill Gap. These ice-contact stratified drift deposits are interbedded with tills and mark recessional ice margins. A major, though discontinuous, recessional moraine can be traced from Old Tappan through Lake Idlewild to Tappan and Sparkill (Fig. 5b; Woodworth, 1905). This moraine is graded to the higher outwash terraces in the Harrington Park area, and is cross-cut by outwash trains from recessional margins to the north. The moraine-outwash sequence is overlain by reddish-brown, laminated, fine sand, silt, and clay of Glacial Lake Hackensack (Averill and others, 1980). The Lake Hackensack deposits can be traced from DeForest Lake, NY, to Elizabeth, NJ (Lovegreen, 1974; Averill and others, 1980).

The outwash trains and the Sparkill Moraine are cut by light brown to gray, trough-cross-bedded, gravelly sand. The sand is reverse-graded locally (see figs. 3 and 6, stops 13 and 15 in Averill and others, 1980). The gravelly sand underlies a 50-foot (15.3 m) terrace and a 30-foot (9.1 m) channel at Sparkill (Fig. 5b). The terrace falls to 30 ft (9.1 m) at Harrington, NJ. The channel falls to 25 ft (7.6 m) at Harrington, where it merges with the terrace and the "fluvial sands" of Lovegreen (1974).
Figure 9. STOPS
The terrace-channel-"fluvial sand" deposit forms a ribbon that disconformably overlies the Hackensack clay from Sparkill, NY, to Elizabeth, NJ (Lovegreen, 1974). They overlie a scoured surface that is carved into the dissicated, oxidized Lake Hackensack clay fill of the Oradell Channel (figs. 15 to 18 in Lovegreen, 1974). They grade into a 15 ft (4.6 m) ASL channel fill at Elizabeth, NJ (fig. 15 in Lovegreen, 1974). The channel deposits grade into reddish-brown, massive to planar-bedded silty sand to sandy silt north of Sparkill.

Stop 1A will be at Tallman Mt. State Park on the 190-foot (58 m) strath terrace of the Sparkill Gap, overlooking the outlet of Lake Albany. Stop 1B will be at the center of the 30-ft channel on Oak Tree Road (Fig. 5b).

STOP 2- Dutchess Quarry Cave: The archeological sites at Dutchess Quarry Cave are on the west flank of Mt. Lookout, a knob of calcitic dolostone that overlooks the Black Dirt region on the Warwick 7-1/2 minute quadrangle of Orange County, NY. The Black Dirt region is an extensive deposit of Holocene fresh-water peat that formed over the Pleistocene 400-foot (120 m) Glacial Lake Wallkill clays (Connally and Sirkin, 1967, 1970). The site consists of rockshelters that contain a major and very old assemblage of vertebrate fossils and human artifacts. The following report by Robert Funk and David Steadman summarizes the important points of the site:

Paleontology and Archeology of Dutchess Quarry Caves, Orange County, New York

David W. Steadman and Robert E. Funk, New York State Museum, The State Education Department, Albany, NY 12230

Much of this account is taken from our paper in Current Research in the Pleistocene, vol. 4, pp 118-120 (1987). The Dutchess Quarry Caves are located in the town of Goshen, Orange County, NY (lat. 41° 22' N, long. 74° 22' W), at an elevation of 177 m on the northwest side of Mount Lookout. The Dutchess Quarry Cave complex consists of a large cave (no. 1; hereafter DQC 1) and lesser caves and fissures (nos. 2-9) that have formed in calcitic dolostone of the Halcyon Lake Formation. Excavations at DQC 1 in the late 1960's yielded several projectile points of the Woodland and Archaic stages from the upper deposits (strata 1A and 1B), one point of the Archaic stage from the top of underlying stratum 2, and a fluted point similar to the Cumberland-type from the base of stratum 2. Associated with these artifacts were bones of 44 species of fish, amphibians, reptiles, birds, and mammals. Of these vertebrates, only the caribou (Rangifer tarandus), reported from both strata 1 and 2, but probably originally entirely from stratum 2, had not been
recorded historically from New York. A radiocarbon date of 12,510 ± 370 years BP (I-4137) was determined on collagen in the caribou bones from stratum 2.

This date may not be a valid age determination for the occupation of Orange County by Paleo-Indians because of ambiguity concerning the association of the caribou bones with the fluted point in stratum 2. Also, the date is about 1,000 years older than the oldest accepted dates for the Clovis fluted point tradition of the Great Plains and Southwest. The point from DQC 1 is most similar to late Paleo-Indian styles, and is probably no older than 10,500 years. Even the Clovis-like points of the Northeast have been radiocarbon dated to an average age of 10,500 years, with a maximum of about 11,000 years.

In 1978 and 1979, J. S. Kopper of C. W. Post College used a resistivity meter to locate eight more cavities in the dolostone of Mount Lookout. The most significant of these was Dutchess Quarry Cave no. 8 (hereafter DQC 8), a much smaller cave than DQC 1 and located about 15 m east of it. Excavations revealed five major strata in DQC 8, with Archaic projectile points in stratum 3, two fluted points at the base of stratum 3, and one fluted point in stratum 5. A radiocarbon date of 5,880 ± 340 years BP (Dic-1447) on charcoal from stratum 3 apparently refers to a Late Archaic occupation. Kopper died in 1984. Although most of the lithic artifacts from DQC 8 had been deposited in the New York State Museum, the charcoal, sediment samples, and bones were obtained only in February 1987 and are now under study. Detailed field records are still missing.

We visited DQC 8 in November 1986. Nearly all of the sediment had been removed by Kopper. Cemented by flowstone to the walls of DQC 8 were remnants of a bone-breccia. Using a hammer and chisel, we collected 11 samples of the bone-breccia, each c. 1 liter in volume. This breccia occurred at all depths from the original surface of the deposit to the deepest excavated level. Probably only the lowest portions of this breccia can be correlated with the breccia zone recorded by Kopper near the cave mouth between strata 3 and 4. The recently collected breccia samples have been etched with acetic acid, revealing bits of charcoal and hundreds of very small bones and bone fragments in each sample. A few plant macrofossils and crude, undiagnostic chert flakes also have been found in this breccia.

Thus far we have identified 45 species of vertebrates from DQC 8, from the recently collected samples as well as those of Kopper. Extinct species include Castoroides ohioensis (giant beaver) and Tayassuidae sp. (peccary).
Plans for the Dutchess Quarry Cave complex include the identification of bones and plant macrofossils, the radiocarbon dating of charcoal from Kopper's excavations and the recently collected breccia samples, additional field reconnaissance of the surrounding area, and further excavation. We hope to obtain a better understanding of relationships among the stratigraphy, chronology, cultural remains, flora, and fauna of this important site.

STOP 3- DUTCHESS COUNTY SOUTH: This stop will include discussions by Roger Case of the USDA-SCS on the soils of the southern Hudson Valley, and by Eric Hanson on the till-over-lacustrine deposits south of Poughkeepsie. We will also discuss the radiocarbon dated log from a foundation excavation for IBM in the Poughkeepsie area that might date the fall in water level from the Lake Albany to the Lake Quaker Springs stage.

The ground water studies conducted in the Casper Creek area by Eric Hanson have revealed abundant evidence for a readvance into Lake Albany. Casper Creek is a tributary to the Hudson River that lies on the Poughkeepsie and Wappingers Falls 7-1/2 minute quadrangles. It enters the Hudson River near the Dutchess Stone Quarry, south of the City of Poughkeepsie. The creek flows over a deep, buried valley. A typical section, developed from test borings in the region near the mouth of Casper Creek, is as follows:

- Surface to 5 ft (1.5 m): lacustrine silt
- 5 to 30 ft (1.5 to 9.1 m): clay
- 30 to 35 ft (9.1 to 10.7 m): till with many rounded clasts, thins to the south
- 35 to 50 ft (10.7 to 15.2 m): clay
- 50 to 52 ft (15.2 to 15.8 m): gravel
- 52 to 60 ft (15.8 to 18.3 m): till with both rounded and angular clasts
- circa 60 ft (18.3 m): shale bedrock

The basal till resembles the upper till except for a higher proportion of angular stones. The upper till has not been detected in the Wappingers Creek Valley, but it has been encountered in the borings for the Newburgh-Beacon and Mid-Hudson Poughkeepsie bridges (New York State Department of Transportation) and the Sprout Creek-Fish Creek valleys (Moore, 1982). It extends from the Poughkeepsie moraine (Connally and Sirkin, 1986) 10 miles (16 km) south to Newburgh. We do not have sufficient data to trace the upper till as far south as the northern edge of the Hudson Highlands. The lower till might be pre-Woodfordian in age, although it appears to be unweathered. The upper till could be either lodgement till of the Woodfordian Glacial Advance or a readvance till.

STOP 4- WEST ATHENS HILL: West Athens Hill is a chert quarry and workshop that was used from Paleo-Indian to Colonial times. The
site is on a ridge of the Mount Merino chert member of the Ordovician Normanskill Formation, and is on the Hudson North 7-1/2 minute quadrangle. The ridge is a remnant of the Late Cenozoic Albany Peneplain (Dineen, 1987). The site overlooks a plain of Lake Albany clay and silt. Test borings on the plain encountered over 160 ft (50 m) of clay and silt. The following report is by Robert Funk:

Archeology of the West Athens Hill Site
Robert Funk, New York State Museum

The West Athens Hill archeological site occupies the summit of a ridge of Normanskill Shale in Greene County, NY, near the Town of Catskill. Paleo-Indians were attracted to this site approximately 11,000 years ago by outcrops of gray to green Normanskill (Mount Merino) chert that provided excellent raw material for their characteristic fluted points and other chipped stone tools. The site served as a quarry-workshop and temporary camp. Major excavations by the New York State Museum in 1966, 1969, and 1970 recovered over 1,500 artifacts including 11 fluted points and the probable fragments of 38 others. Unfortunately, there were no datable organics from the Paleo-Indian occupation, and osseous remains of humans and food animals were absent due to highly acid soil conditions. The hill is located within the bounds of Glacial Lake Albany, therefore the earliest human occupation post-dated the draining of the lake.

STOP 5- POWELL-MINNOCK CLAYPIT: The Powell-Minock claypit is on the Ravena 7-1/2 minute quadrangle in Coeymans, Albany County, NY. The pit supplied the Powell-Minnock brickworks from the turn of the century until the nineteen fifties. Powell-Minnock presently make their bricks from the Esopus Shale.

The claypit is in a clay plug that fills the mouth of the buried Colonie Channel. The Colonie meets the preglacial Battenkill-Hudson under the tidal flats of the present Hudson River. The floor of the Colonie Channel beneath the pit lies at an elevation of 110 ft (33.3 m) below sea level (Dineen and others, 1983). A composite stratigraphic section has been measured at the pit, a few grain-size analyses have been run (Gail Ashley at Rutgers), and measurements of paleomagnetism have been performed (William Brennen, SUNY at Geneseo). Gail Ashley noted the presence of insect larvae tracks in the silt rhythmites at this site. The lithic section is as follows:

(Top) 19.4 to 16.4 m (62.3 to 53.8 ft): 10YR 4/2 clay and 10YR 8/2 fine silt rhythmites, 1 to 10 cm (0.5 to 4 in) thick, 5 cm (2 in) average. Rhythmites are planar, with some convoluted and flame structures. Flames are overturned to the west. The sequence fines upward and is 50% clay. "Clay
dogs" or doughnut-shaped concretions are present. (WEATHERED DEEP-WATER FACIES)

16.4 to 14.9 m (53.8 to 48.9 ft): 10YR N/5 rhythmic clay with less than 5% silt. Rhythmites are 0.5 to 10 cm (0.2 to 4 in) thick, thinning upward. One ripple-laminated, fine sand bed occurs under the top rhythmite. The size of the clay is 8 phi. Insect larvae tracks were found in this layer. (DEEP-WATER FACIES)

14.9 to 12.3 m (48.9 to 40.4 ft): 10YR 2/2 mottled clay and fine-sandy silt in 1 to 10 cm (0.5 to 4 in) thick rhythmites. The rhythmites thicken and become coarser upwards. The silty members are climbing ripple-laminated, the ripples climb to the N90E and are capped with fine sand. (DISTAL TURBIDITES)

12.3 to 11.4 m (40.4 to 37.4 ft): 10YR 5/2 contorted, fine to medium sand, with some 1 to 2 cm (0.5 to 1 in) laminae of clay. The sand contains abundant flame structures that are overturned to the west. Blebs of clay are abundant at the base of this bed. (TURBIDITE)

11.4 to 9.7 m (37.4 to 31.8 ft): 10YR N/4 massive to poorly layered clay. Laminae are overturned to the S80W, the top of the bed has ridges that trend N57E to N10E. (TURBIDITE)

9.7 to 9.1 m (31.8 to 30 ft): 10YR N/4 rhythmic clay and silt, with dropstones near the top of the unit. The dropstones are as large as 2cm by 2cm by 8cm (1 by 1 by 3 in) and are rounded. Rhythmites are 4 to 15 cm (2 to 6 in) thick, and are 50% clay. (DEEP-WATER FACIES)

9.1 to 7.5 m (30 to 24.6 ft): 10YR N/4 massive to contorted clay with 5 to 10% silt and 50 cm (20 in) thick fine sand beds. (TURBIDITE)

7.5 to 4.1 m (24.6 to 13.5 ft): 2.5Y 7/2 silt and 10YR N/4 clay and silt rhythmtes. Silt is ripple-laminated to cross-bedded. Some 2 to 5 cm (1 to 2 in) rounded pebbles occur in the clay. One 100cm by 100cm by 200cm (40 by 40 by 80 in) striated graywacke boulder or dropstone was present. The bedding is deformed under the large dropstone. The rhythmtes are 8 to 35 cm (6 to 12 in) thick, decreasing in thickness and silt content upwards, they dip towards the southwest (dips range from S20E to N70W). Some vertical N15E joints are present. (ICE-CONTACT FACIES)

4.1 to 2.8 m (13.5 to 9.2 ft): 10YR 5/2 very fine to medium sand, with 25N60W climbing ripples at the base, N60W ripple-trough-laminae in the center, and 20S80W climbing ripples at the top. This unit fines upward. (ICE-CONTACT FACIES)
2.8 to 2.6 m (9.2 to 8.5 ft): 5Y 5/3 to 5Y 6/3 fining upward, ripple-laminated sand, silt, and clay. (ICE-CONTACT FACIES)

2.6 m to base: 10YR 5/2 cross-bedded cobbly sand and gravel. Beds are 10 to 15 cm (4 to 6 in) thick and fine upwards. Some imbricated cobble beds, with a silty sand matrix. Cobbles are rounded and range in size from 2 to 12 cm (1 to 5 in). The cross-beds dip from S40W to S10W. The sequence is faulted from the base to 4.1 m (13.5 ft). These normal faults trend S50W and dip 35S. (ICE-CONTACT FACIES)

Test borings in the claypit suggest that the ice-contact sand and gravel extends to a depth of 170 ft (52 m), that it overlies 5 to 10 ft (1.5 to 3 m) of till, and that its top is quite uneven.

STOP 6- MASICK (ROTTERDAM) PIT: The Masick sand and gravel pit lies in the Rotterdam Junction 7-1/2 minute, quadrangle in Schenectady County, at an elevation of 340 ft (104 m). The pit is in the southwest corner of the Schenectady Delta, at the point where the Normans Kill entered Lake Albany.

The pit exposes 10 to 20 ft (3 to 6 m) of 10YR 5/4 planar cross-bedded, cobbly sand and gravel at its base, the beds dip towards the southeast. They are overlain by 5 ft (1.5 m) of 10YR 5/4 trough-cross-bedded sand and gravel, the trough axes trend east-west. The tops of the sand and gravel units are sheared, and overlap by 1.5 to 10 ft (0.5 to 3 m) of 10YR 4/4 compact, fissile, matrix-supported, bouldery, sandy, clay diamicton. The diamicton is overlain by 1.5 to 13 ft (0.5 to 4 m) of 10YR 4/3 climbing ripple-laminated (base), to planar and ripple-trough-laminated (center), to climbing ripple-laminated (top) coarse to fine sand, fining upward. The ripples climb S0E to S50E. The sand beds are overlain by 13 ft (4 m) of trough-bedded, gravelly sand. The troughs are 8 to 12 in (20 to 30 cm) deep and 5 to 10 ft (1.5 to 3 m) wide, and trend southeast. The top of the sequence is 6 to 10 ft (2 to 3 m) of 10YR 4/3 planar to trough-laminated fine to medium sand.

The basal sands and gravels are a kame delta associated with a recessional ice margin at Schenectady. The ice front readvanced across the kame delta, and deposited the gray, compact till. The glacial Normans Kill deposited the upper gravelly sands over the abandoned, till-covered kame after the ice retreated. Lake Albany did not fall to the Lake Albany II stage until after the 340-foot (109 m) Normans Kill and Schenectady delta was deposited.

STOP 7- WUNDERLICH (Pollock Road) PIT: The Wunderlich pit was Stop 1 in Dineen and Rogers (1979). It is in the Pollock Road kame delta, in northeastern Albany County, on the Niskayuna 7-1/2 minute quadrangle. Meltwater from the Niskayuna ice margin
deposited a network of eskers from Clifton Park to Pollock Road (Dineen and others, 1983). These eskers grade into the northern edge of the Pollock Road kame delta. The delta was deposited in Lake Albany at the Niskayuna ice margin. It was abandoned, and the ice front had retreated out of the area, by the time the lake level fell to the Albany II stage.

The pit exposes 20 ft (6 m) of 10YR 6/3 normal faulted, trough-cross-bedded sand and gravel. The beds dip to the southwest, the faults are down to the north. These ice-contact deposits are overlain by 65 ft (20 m) of 10YR 5/4 gravity and thrust-faulted, planar cross-bedded gravelly sand, with contorted beds near the top of the unit. The cross-beds dip to the south. The top of the section is a 3 to 20 ft (1 to 6 m) 10YR 6/4 matrix-supported, bouldery, silty sand diamicton. The central beds grade into rhythmically laminated silt and clay to the south. The upper till does not extend to the south.

The sequence in the Wunderlich pit is very similar to that in the Masick Pit. Their deposits are probably contemporaneous and record the same minor readvance.

STOP 8- LUNCH, GOOD TIMES RESTAURANT, ALPLAUS CHANNEL: The present landscape of the Hudson Valley between Schenectady and Saratoga Springs includes many large valleys that contain underfit streams and lakes. The valleys were carved by water flowing out of the Mohawk Valley and into Lakes Quaker Springs and Coveville.

The Schenectady delta was deposited as the Hudson Lobe retreated from the Meadowdale (Fig. 7b) to the Niskayuna margins (Hanson, 1977). Lake Albany clay filled the preglacial Alplaus-Ballston channel as the delta grew. The delta continued to grow, and the animals at Glenville continued to thrive, as the glacier retreated to the Hidden Valley-Glen Lake-Carter-Manchester ice margin (Fig. 7c). The lake level fell to the Albany II stage and the delta was abandoned. The emergent delta deflected the Mohawk River flow into the Alplaus-Ballston channel.

Catastrophic floods from the Ontario Basin started to carve out the clay in the Alplaus-Ballston channel, and to scour a series of overflow channels across the interfluve between the preglacial Alplaus-Ballston and Colonie channels (Hanson, 1977; Dineen and others, 1983). Lobes of sand and gravel were deposited in Lake Albany II near Shenendahowa and Clifton Park (Dineen and others, 1983). Slackwater deposits from later floods backfilled the interfluve overflow channels. Most of the flood waters passed through the East Line, Drummond, and Ballston Creek channels into Lakes Quaker Springs and Coveville. A bedrock sill developed at East Line between the Drummond and Ballston Creek channels. The overflow channel gravels are a source of groundwater in southern Saratoga County.
The Drummond channel carried the floods across the present site of Saratoga Lake and formed the Schuylerville plunge pool. At the same time, flood overflows deepened the Ballston Creek channel. The Ballson Creek channel eventually "pirated" flood flow into and through the Round Lake basin. By the Fort Ann stage, the overflows from the catastrophic floods had reduced the floor of the Rexford channel (south, near Schenectady) enough to divert the the Mohawk River into its present course.

A controversy is presently raging between Dineen and Hanson concerning the origin of Ballston, Round, and Saratoga Lakes. Hanson cites abundant evidence for the existence of recessional ice margins at Niskayuna, Round Lake, and Saratoga Lake (Hanson, 1977). He suggests that thick blocks of stagnant ice were buried under the fluvial deposits at these margins. The lakes are in kettle holes left by the melted ice blocks.

Dineen points out that Round and Saratoga Lakes overlie the intersections of the Drummond and Ballston Creek channels with the preglacial Colonie channel (Dineen and others, 1983). He suggests that the lakes occupy large scour holes carved into the relatively soft glacial deposits that fill the Colonie channel. He also suggests that Ballston Lake also lies in a deep scour hole.

STOP 9- CHAMPLAIN CANAL LOCK 1, CAMPBELL ISLAND: This exposure is in Rensselaer County, on the Troy North 7-1/2 minute quadrangle. It is on the bank of a little "cove" or widening of the Hudson River. The clay and silt deposits exposed here fill a section of the Battenkill-Hudson Channel that lies east of the present Hudson.

Exposures of Lakes Albany and Quaker Springs rhythmites lie along the riverside road for a mile. The road has been closed for several years because of large landslides that occurred after heavy seasonal rains hit during the spring thaw. The deluge coincided with the inability of the Town of Schagticoke to pay for repairs. Observations in the Albany area have shown that the clays are potentially unstable if the surface slopes exceed 12 degrees, especially after heavy rains or if the slopes have been undercut or loaded.

The continuous horizontal laminae of clay, silt, and fine sand are occasionally interrupted lenses or pods of pebbly clay diamicton that range from a few centimeters to half a meter. The clays are typically contorted or disturbed under the diamicton lenses.

The compact, pebbly, unsorted nature of the diamicton pods and the deformation of the underlying sediments indicate that these pods and lenses settled through the water column and impacted the
soft, oozy lake bottom. Hence, these diamictons are dropstone tills, presumably derived from icebergs and iceshelves drifting in the prevailing currents of Lakes Albany and Quaker Springs. The stratigraphic position of these exposures, along with their northern location, suggest that these deposits date from the late Lake Albany and early Quaker Springs stages. The numerous dropstone till lenses and pods along discrete horizons in the rhythms of the Deep-water Facies represent intervals of accelerated ice margin retreat by calving. We do not have enough information to determine whether calving was a major process in the retreat of the Hudson Lobe.

The water was at least 100 ft (30 m) deep along the major axes of the deeper north-south basins of the lake. The gamma-ray and resistivity stratigraphy developed in Dineen and others (1983) suggest that "grounding lines" developed along the ice front in areas where the gradients of the preglacial channels changed abruptly. Relatively stable ice margins formed at those grounding lines.

STOP 10- FORT ANN OUTLET: The rock outcrops along US Route 4 on the Fort Ann 7-1/2 minute quadrangle exhibit numerous potholes, particularly in the constriction of the primary Fort Edward channel adjacent to the Champlain Canal. Chapman (1937) interpreted these potholes as the product of major stream discharge from Lake Fort Ann, in the Champlain Valley, into the Hudson Valley. Three distinct Fort Ann levels are documented by a series of terraces, some in the outlet channels, that have been mapped in the Fort Ann area (DeSimone, 1985; DeSimone and LaFleur, 1986). The water planes (DeSimone and LaFleur, 1986) and the presence of the potholes suggest that the Fort Ann stage was fluvial in the Hudson Valley.

The Fort Ann water levels can be correlated with the water levels in the St. Lawrence and Ontario basins (DeSimone and LaFleur, 1986). Pulses of high discharge and rapid water level changes in the St. Lawrence and Ontario Lowlands (Clark and Karrow, 1984) coincide with periods of enhanced erosion in the Fort Ann outlet channels (DeSimone and LaFleur, 1985, 1986). The secondary Durkeetown and Winchell channels (DeSimone and LaFleur, 1986) might have been used as overflow routes for part of the enhanced discharge. The stable level of Fort Ann fell after each pulse of increased discharge.

STOP 11- LODGEMENT TILL AND BASCOM CLAY: We are now travelling into the Taconic Mountains section of the New England Highlands, and are on the North Pownal, VT-NY, 7-1/2 minute quadrangle. The roadside exposure is part of a borrow pit. The exposure is 8 to 10 ft (2.4 to 3 m) high, and consists of 6 to 8 ft (1.8 to 2.4 m) of tan to gray silt and clay rhythms over gray, compact, silty-clay, matrix-supported diamicton. The diamicton contains abundant pebbles, some cobbles, and a few boulders that are
predominently local gray shales, slates, and phyllites with some quartzite and marble. The stones are subrounded to rounded, with striations and crescentic marks. The diamicton is the typical lodgement till found in the Taconics and the Hudson Valley. It sometimes directly overlies polished, striated rock pavements. Its texture and composition reflect the fine-grained texture of the local bedrock.

The rhythmites represent Glacial Lake Bascom clay. The upper, tan clay is oxidized. Fresh exposures reveal alternating dark gray and light gray couplets.

Drill holes at Williamstown, MA, reveal 90 ft (27 m) of lake clay and silt under the Hoosic River floodplain. The general lacustrine stratigraphy near Williamstown parallels the sequence of lake sediments in the Hudson Valley discussed previously. A basal, coarse-grained, Ice-contact Facies representing eskers and subaqueous fans grades up to a thick silt and clay, Deep-water Facies. A truncation surface separates the lake sediments from the overlying late Pleistocene and Holocene fluvial deposits of the Hoosic River. The basal Deep-water Facies is a confined aquifer that is probably recharged from the valley-side kame deposits. The fluvial sand and gravel deposits are a shallow aquifer that is not hydraulically connected to the deep aquifer.

Upland areas with appreciable lake deposits that consist of more than a few isolated deltas demonstrate the presence of a persistent glacial lake in an upland drainage basin. Mappers in upland areas should look for extensive clay deposits to test whether isolated deltas are small ice-marginal lake features or are part of a major lake sequence.

STOP 12- LAKE BASCOM FAN DELTA, MORE SAND AND GRAVEL PIT: The More Sand and Gravel Pit is on the North Pownal, VT-NY, 7-1/2 minute quadrangle. The main face of the excavation exposes 10 to 15 ft (3 to 4.6 m) of poorly-sorted, horizontally bedded to massive cobble, pebble, and small boulder gravel. The clasts in this sheet gravel are predominantly local Taconic phyllite, with only minor quartzite and oxidized, reddish-brown cobbles of a pre-Woodfordian (?) regolith. Similar clasts of oxidized regolith were observed in the Guilderland area west of Albany (Dineen and Rogers, 1979). The regolith clasts can easily be found because of their bright color and low density. The phyllite and quartzite clasts are angular to subrounded, and the regolith clasts are subrounded to round. A finer-grained basal facies of interbedded sands, pebble gravels, and silts that dips towards the valley axis was exposed in a test pit in the floor of the excavation. The sheet gravel is separated from the underlying finer deposits by a truncation surface. Well-sorted, laminated, fine sands and ripple-laminated sand are exposed up the hill from the gravel
pit. These fine sands are capped by a thin (1 to 3 ft, 0.3 to 1 m) pebble gravel. The morphology of the overall deposit suggests that it was deposited in the ice-contact environment.

As the Hoosic Valley ice tongue retreated from the North Pownal area, it deposited subaqueous fans of sand and gravel into Glacial Lake Bascom. The Potter Hill spillway (on the Grafton, NY, 7-1/2 minute quadrangle) was opened up as the glacier retreated towards North Petersburg, NY. Lake Bascom fell to the 900-foot (274 m) level, and the 1,005-foot (306 m) Berkshire outlet (north of Pittsfield on the Pittsfield, MA, 7-1/2 minute quadrangle) was abandoned. See Taylor (1903) for more information on early Lake Bascom, and DeSimone and Delthier (1986) for late Lake Bascom's history.

Several deposits are graded to the 900-foot (274 m) level of Glacial Lake Bascom. These include the Green River–Roaring Brook outwash at South Williamstown (Williamstown, MA 7-1/2 minute quadrangle), the Little Hoosic River outwash in Center Berlin, and the Satterlee Hollow fan delta on the Berlin, NY–MA, 7-1/2 minute quadrangle.

Lower Lake Bascom spillways were exposed as the glacier retreated from the North Petersburg area. The 700-foot (213 m) Lake Bascom was controlled by the Otter Creek meltwater channels that head at Nipmoose Hill (Eagle Bridge, NY, 7-1/2 minute quadrangle). The 700-foot (213 m) level was short-lived, and is recorded by the 700-foot (213 m) Petersburg terrace, the Green River fan near Sweets Corners (Williamstown, MA, 7-1/2 quadrangle), and the 700-foot (213 m) terrace west of Pownal.

The 625-foot (191 m) Case Brook col near Hoosick Junction (Hoosick Falls, NY, 7-1/2 minute quadrangle) was exposed as the Hudson Lobe withdrew to the Niskayuna ice margin (DeSimone and LaFleur, 1986). This allowed the level of Lake Bascom to fall again, to 620 to 630 ft (189 to 192 m). The More Pit is in an alluvial fan delta deposited into this lake stage as meteoric waters eroded the unvegetated tributary valleys. Other fan deltas occur at the mouths of Ladd Brook, Dell Creek, Church, North Church, Potter, Reservoir, Frost, Lincoln, and Ellis Mine Hollows in the Hoosic and Little Hoosic Valleys.

The water level fell to 520–540 feet (158–165 m) as the ice front reached the Waterford ice margin near North Hoosic (DeSimone and LaFleur, 1986). The water probably escaped down the Hoosic Valley to Lake Albany along the thinning ice margin. Alluvial fan deltas were deposited at North Petersburg, Hoosic, and Hoosick Junction.

Terraces in the Hoosic Valley at 450–460 and 410–420 feet (137–140, 125–128 m) suggest that Lake Bascom fell to still lower levels with further retreat of the ice. As the ice retreated to the Willow Glen ice margin (DeSimone and LaFleur, 1986), the
early, 370-foot (113 m) delta of the Hoosic River was deposited in Lake Albany.

STOP 13- FISH HATCHERY KAME: The last formal stop is on the Pownal, VT, 7-1/2 minute quadrangle. We will drive through the backroads so that we can develop an appreciation for the subtle distinctions between the streamlined, active ice landscapes and the hummocky areas of ice disintegration. The smoothly rounded drumlin and drumlinoid forms of the streamlined landscape express the imprint of ice abrasion and till deposition. The more chaotic, rolling ice disintegration landscape shows the effects of meltwater erosion and deposition under and in contact with rotting ice. The kames, kettles, and eskers can be variously mapped as kamefields, kame stagnation moraine, or undifferentiated kamic topography, depending on the individual preferences of the mapper. When you distinguish the landforms, you make a basic materials and process evaluation as well.

The Fish Hatchery Kame exhibits 35 to 40 ft (11 to 12 m) of west and southward-dipping, interbedded and cross-bedded, moderately-to well-sorted, faulted and folded, gravel and sand. The deposit contains an unusually high proportion of black slate and phyllite grains. It contains 70% rounded tan to white quartzite grains, 20% rounded and flattened black slate and phyllite grains, and 10% rounded, flattened green phyllite, slate, and shale grains. The dark hue of the deposit is caused by the high percentage of dark lithic grains.

The deposit is capped with a 2 to 4 ft (0.6 to 1.2 m) thick cobbly, pebbly, sandy, matrix-supported diamicton. The till is loosely compacted and is oxidized to brown. The major esker in the valley heads at this stop, and fed a small delta in an ice-marginal lake. The lake formed between the previous ice margin, southwest of Pownal Center, and the delta. The lake flooded the lower flanks of the Dome and the Green Mountains (DeSimone and Delthier, 1987).

The deposit is the product of nearby active ice that supplied sediment scoured from the outcrops of Hortonville Slate a few miles to the north. Longer transport would have destroyed the soft slate grains. The slate grains do not appear in exposures to the south. The upper diamicton is a weathered meltout till. Depth-to-bedrock studies in the Bennington area show a thick wedge of valley fill, and strengthen the interpretation that the ice to the south was stagnant while active ice occupied the valley to the north. The distribution of the deposit support the Malaspina Glacier analogue of Gustavson and Boothroyd (1987) for deglaciation in New England. The deposits throughout the Hoosic Valley consist of kames and eskers that contain stratified meltwater deposits capped with meltout till, suggesting subglacial deposition under rotting ice.
The predominance of ice stagnation features and the observed sediment textures and stratigraphy in numerous pits and exposures suggest that Glacial Lake Bascom did not invade this section of the Vermont Valley, as recently outlined by Warren and Stone (1986). Warren and Stone (1986) perpetuated the errors of previous workers whose illustrations of Glacial Lake Bascom carried the +1,000-foot (305 m) (1,060 to 1,080 ft, 323 to 329 m, in the Pownal area) waters into the Vermont Valley (see figure 9 in Bierman and Dethier, 1986). Detailed mapping (DeSimone and Dethier, 1987) and ice margin correlations (DeSimone and LaFleur, 1986) indicate Lake Bascom fell to the 900-foot (274 m) level controlled by the Potter Hill spillway while stagnant ice choked the southern part of the Vermont Valley. Lake Bascom sediments are absent in that part of the valley. North of Bennington, lake silt and clay is present and has been attributed to Glacial Lake Shaftsbury (Shilts, 1966; Stewart and MacClintock, 1969). Shilts (1966) and the reconnaissance-level map of the Bennington 15 minute quadrangle do not adequately indicate the limit of Lake Shaftsbury. Lake Shaftsbury was probably coeval with the 900-foot (274 m) stage of Lake Bascom. DeSimone plans to continue detailed mapping on the Bennington 7-1/2 minute quadrangle to settle this question.
STRATIGRAPHY AND PROVENANCE OF DIAMICTONS AND OUTWASH DEPOSITED BY THE SOUTH CHANNEL LOBE IN EASTHAM AND WELLFLEET, CAPE COD, MASSACHUSETTS

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INTRODUCTION

This field trip will focus on the stratigraphy, sedimentology, and provenance of diamicton and outwash that occurs in the Eastham and Wellfleet plains (Fig. 2). Considerable attention will be paid to the lithology and mode of deposition of the diamicton facies that occur in these outwash plains. The diamictons have heretofore received only cursory study. The diamicton facies at the study sites are overlain by sandy outwash; respectively, these materials are inferred to record deposition on a subaqueous delta or fan that was succeeded by a braid delta. Deposition of the diamicton facies was by subaquatic glaciogenic debris flows, whereas interbedded sorted sands and silts are products of turbidity currents. The sandy outwash is believed to reflect deposition by braided streams and subaqueous deltaic outwash cones abutting glacial Lake Cape Cod. The source of the diamicton and outwash was the South Channel lobe.

The Pleistocene history of eastern Massachusetts has been the subject of continual study since Conrad (1839) and Hitchcock (1841) initially espoused the applicability of the glacial theory to the New England area. The Cape and islands region of eastern Massachusetts has been of particular interest, owing to complex interrelationships among the three lobes of the late Wisconsinan ice sheet which were active during deglaciation of the region (Fig. 1). From west to east, they are named the Narragansett Bay-Buzzards Bay lobe (hereafter called simply the Buzzards Bay lobe), the Cape Cod Bay lobe and the South Channel lobe. Retreat of these lobes apparently was not synchronous, as various stratigraphic evidences show the deposits of the Buzzards Bay lobe to be older than the deposits of the Cape Cod Bay lobe (Mather et al., 1942), which are in turn older than the deposits of the South Channel lobe (Oldale,
Figure 1. (A) Regional geology of New England and Cape Cod region (after Ballard and Uchupi, 1975). Dashed line shows generalized path of South Channel lobe toward Eastham and Wellfleet (B), based on glacial erratics, geomorphic and stratigraphic considerations. (B) Cape Cod region showing position of stagnating ice margins (sawteeth to north or east) and advancing ice margins (sawteeth to south) at time of deposition of Sandwich Moraine (SM), Wellfleet Plain (WF), Truro Plain (T), and Eastham Plain (EP) (after Oldale, 1982). Inset map shows generalized configuration of glacial lobes during retreat of the eastern part of the Laurentide ice sheet: NBL - Narragansett Bay-Buzzard's Bay lobe; CBL - Cape Cod Bay lobe; SCL - South Channel lobe.
Figure 2. Outwash plains of the South Channel lobe on outer Cape Cod. Contours are drawn on the uncollapsed surfaces of the Eastham and Wellfleet Plains. Contour interval ten feet. (after Oldale, 1968, Oldale and Stone, 1987; Oldale et al., 1971).
This differential retreat from west to east resulted in the establishment of large proglacial lakes: Glacial Lake Taunton abutting the Buzzards Bay lobe and Glacial Lake Cape Cod in front of the Cape Cod Bay lobe (Larson, 1982).

The existence of a proglacial lake in Cape Cod Bay was first proposed by Grabau (1897), who believed that the outwash plains on outer Cape Cod (Truro, Wellfleet and Eastham areas) were deltas which had prograded westward into such a lake (Fig. 2). Woodworth and Wigglesworth (1934) suggested that the deltas were fed by not only the Cape Cod Bay lobe, but also by a lobe whose terminus lay immediately east of Cape Cod, which they named the South Channel lobe. Glacial Lake Cape Cod was confined to the south by the Sandwich Moraine, and to the east by the South Channel lobe, which was also a major source of meltwater (Figs. 1, 2; Larson, 1982; Oldale, 1982).

These earlier efforts have since been substantiated by the detailed geologic quadrangle mapping on Cape Cod by Oldale and co-workers (summarized by Oldale and Barlow, 1986). Their results very clearly demonstrated the westward-sloping nature of the outwash plains on outer Cape Cod, and hence their genetic relationship with the South Channel lobe. Furthermore, the pebble provenance of diamictons and sandy outwash implies that the South Channel lobe eroded and incorporated debris from bedrock similar to that found presently in eastern New England (Fig. 3, and discussion below).

STOP 1: Coast Guard Beach

Coast Guard Beach is reached from Route 6 by driving east on Nauset Road past the Salt Pond Visitor Center. Limited parking is available at the beach itself except during the peak tourist season (July 1 through Labor Day), when
a shuttle bus must be used from the large parking area near the Visitor Center. The exposure to be examined at Coast Guard Beach begins about 50 m north of the limit of the guarded beach. The elevation of the uncollapsed surface of the Eastham plain at this site is approximately 60 ft. (18 m) (Fig. 2).

The outcrop exhibits a coarsening-upward sequence of massive and stratified diamictons (Figs. 4, 5). Massive, matrix supported diamicton occurs at the base of the sequence. This facies also includes some faintly stratified diamicton whose layering is defined by thin (2-3 mm) stringers of silt and sand. The massive diamicton contains fewer, and smaller pebbles, than the well-stratified diamicton above (Fig. 5). The massive diamicton is conformably overlain by well-stratified diamicton, which is typified by 5-10 cm-thick interbeds of sand and silt that are commonly rippled. Stratification is commonly contorted, and in addition, a variety of small isoclinal folds can be observed in the outcrop. These structures show no consistent orientation. The well-stratified diamicton can be divided into upper and lower units based on stone content and maximum stone size, which increase upward (Fig. 5). In contrast to the lower unit, the upper unit commonly contains local segregations of pebbles, cobbles and boulders.

To date, granulometric, clast fabric and pebble provenance studies have been undertaken only in the massive diamicton facies (Figs. 3, 5, 6). The massive diamicton predominantly consists of sand and silt with very little clay (Fig. 6; Table 1). Clast A-axes dip at shallow angles and are arranged in two bimodal clusters strongly aligned northwest to southeast. Clast C-axes are mostly steeply dipping (Fig. 4). Pebbles in the massive diamicton are commonly striated.
Figure 3. Pebble lithologies of outwash and diamicton facies of the Eastham and Wellfleet plains.
Figure 4. (Stop 1) Lithologic map of diamicton facies at Coast Guard Beach, drawn from serial field photographs taken in 1987. Inset shows location of Figure 5.
Figure 5. (Stop 1) Photograph of diamicton facies at Coast Guard Beach: from bottom to top, massive to very faintly stratified diamicton, well-stratified diamicton with few large clasts, and well-stratified diamicton with abundant large clasts. Equal-area, lower hemisphere stereogram shows pebble fabric measured in massive diamicton facies at "X". Staff is calibrated in 30 cm divisions.
Figure 6. Ternary diagram showing granulometric composition of diamicton facies observed in the Eastham and Wellfleet Plains.
TABLE 1

Granulometric composition of diamictons in the Eastham and Wellfleet plains, showing mean and range of percentage sand (2.0-0.063 mm), silt (0.063-0.002 mm), and clay (finer than 0.002 mm)

<table>
<thead>
<tr>
<th></th>
<th>%Sand</th>
<th>%Silt</th>
<th>%Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Coast Guard Beach:</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Massive Diamicton (N = 5)</td>
<td>56</td>
<td>41</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>(54-59)</td>
<td>(39-43)</td>
<td>(3-6)</td>
</tr>
<tr>
<td><strong>Newcomb Hollow Beach:</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>North Section</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Massive Diamicton (N=12)</td>
<td>8</td>
<td>88</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>(4-15)</td>
<td>(80-92)</td>
<td>(2-5)</td>
</tr>
<tr>
<td>Stratified Diamicton (N=10)</td>
<td>17</td>
<td>74</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>(9-29)</td>
<td>(59-83)</td>
<td>(6-13)</td>
</tr>
<tr>
<td><strong>South Section</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Massive Diamicton (N=4)</td>
<td>5</td>
<td>89</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>(4-6)</td>
<td>(87-93)</td>
<td>(3-8)</td>
</tr>
<tr>
<td>Stratified Diamicton (N=5)</td>
<td>15</td>
<td>80</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>(8-19)</td>
<td>(76-85)</td>
<td>(2-8)</td>
</tr>
</tbody>
</table>
The pebble suite of the massive diamicton contains significant quantities of slate and vein quartz with slate inclusions, which are lacking in Eastham plain sandy outwash. Felsic volcanic rock and quartzite are the two other main rock types in the diamicton facies, and also make up a large proportion of the pebble suite in Eastham plain deposits (Fig. 3). The quartzite cobbles commonly contain fossils of the Cambro-Ordovician brachiopod genus *Obolus*. Similar quartzite pebbles have been observed in South Channel lobe sediments elsewhere on outer Cape Cod (Oldale, 1986, oral communication), and also in glacial drift on Martha's Vineyard (Shaler et al., 1899). The onshore sources of these quartzite clasts are Pennsylvanian conglomerates found in the Narragansett Basin of southeastern Massachusetts and Rhode Island (Perkins, 1920).

**Interpretation**

It is appropriate at this juncture to briefly discuss the earlier work of Sayles and Knox (1943), who also studied glaciogenic sediments in the Coast Guard Beach area. Sayles and Knox (1943) described four "tills" at this site. Their uppermost till is now recognized to be an eolian sand containing ventifacts (Oldale, 1982, p. 4). From the descriptions provided by Sayles and Knox of the underlying "tills" and associated stratified sands and silts ("pseudovarves": p. 1585-1586, and their plates 2 and 3), it is likely that they studied a sedimentary sequence that was similar to the present outcrop. Their photographs clearly illustrated the bedded nature of the glaciogenic sediments as a whole. Moreover, in one case Sayles and Knox (their Fig. 1) showed the upper two "tills" occurring in an "anticline," whose axis trended
east to west. Other smaller folds and faults were also mentioned, and all such deformations were ascribed to glacial movements.

In view of the stratified nature of much of the diamicton facies at Coast Guard Beach, and the occurrence of interbeds of rippled sand and silt, it is more reasonable to re-interpret the diamictons as subaquatic glaciogenic debris flow deposits rather than tills. Deposition probably took place in subaqueous deltas or fans (the "anticline" of Sayles and Knox). The folds, faults and contorted bedding observed in the present study and also earlier by Sayles and Knox probably are penecontemporaneous deformations caused by flowage (Evenson et al., 1977) and loading (Stewart, 1987) during deposition.

STOP 2: Wellfleet Municipal Sand Pits

The sand pits at this stop are most easily reached via Bound Brook Island Road, which is a west exit off Route 6 about 1.2 miles (2 km) north of the traffic light marking the principal exit to the town of Wellfleet.

Observations will be made at two sand pits used for road maintenance and landfill cover material. Only one pit was active in 1987, and hence it is designated the active pit. The sand pits are excavated in sandy outwash of the older Wellfleet plain (Oldale, 1968). The elevation of the uncollapsed surface of the Wellfleet plain here is approximately 100 feet (30.5 m) (Fig. 2).

The most striking features in the active pit are outwash cones (Fig. 7). The outwash cones principally consist of sand, however, thin (10-15 cm) interbeds of silty diamicton may also be observed near the top of the cone shown in Fig. 7. Occasionally, ventifacts may be found in the sandy outwash at the active and inactive pits. At the active pit, the axes of the outwash
Figure 7. (Stop 2) Outwash cone in sandy older Wellfleet Plain deposits at Wellfleet municipal sand pits. Section is transverse to axis of cone. Flow was westward, toward the observer. Note faulting in center of cone, to left of shovel.
cones trend east-west, and bedding dips to the west. A small graben occurs in the core of the cone shown in Fig. 7; smaller normal and reverse faults occur within the graben.

At the inactive pit no outwash cones were observed, but stratified silty diamicton is more prevalent. Bedding in the stratified diamicton also dips to the west (Fig. 8A).

Interpretation

The outwash cones observed at the active pit are interpreted to reflect subaqueous deposition by meltwater streams debouching sediment into glacial Lake Cape Cod. Most of the sandy sediment was probably deposited as bedload from underflows on delta foresets (Jopling and Walker, 1968; Gustavson et al., 1975; Rust and Romanelli, 1975; Sharpe, 1987). The stratified silty diamicton is interpreted to have been reworked from the diamicton facies that underlies older Wellfleet plain outwash. Outwash cones like those observed at this stop may form where sediment-laden currents emerge from subaqueous tunnels at glacier margins (Rust and Romanelli, 1975; Sharpe, 1987). It seems evident that the outwash cones at Stop 2 were produced by confined subaqueous currents; however, there is presently insufficient evidence to speculate on the nature of possible conduits. The bedding orientations at the two sand pits give further confirmation to Oldale's (1982) interpretation of an eastern, South Channel lobe source for the Wellfleet plain based on its westward sloping, uncollapsed surface.
Figure 8. Equal-area, lower hemisphere stereograms of (A) poles to stratification in stratified diamicton facies at Wellfleet municipal sand pits; (B) poles to stratification in stratified diamicton facies at Newcomb Hollow Beach north section; (C) pebble A-axes and (D) pebble C-axes in body of pebble-rich diamicton at Newcomb Hollow Beach north section (Figure 9, 15-45 m). Pebble fabric was measured at the 25 m-mark on Fig. 9. Eigenvector ($V_1$) and normalized eigenvalue ($S_1$) obtained from method of numerical analysis of Mark (1973).
STOP 3: Newcomb Hollow Beach (north section)

Newcomb Hollow Beach is reached from Route 6 by proceeding east on either Gull Pond Road or Gross Hill Road, which merges with Ocean View Drive, and then left (farther east) about 1 km to the municipal parking lot at the beach. The north section at Newcomb Hollow Beach begins just south of the parking lot. The south section (Stop 4) consists of a single outcrop 600 m south of the end of the north section (0 m on Fig. 9). The elevation of the uncollapsed surface of the Wellfleet plain at the north and south sections is approximately 130 ft. (40 m) (Fig. 2). In addition to exemplary exposures of South Channel lobe deposits at Newcomb Hollow Beach, there are also excellent examples of beach cusps and rhythmic shoreline topography (Dolan et al., 1974).

The gross stratigraphy at the north section consists of a diamicton facies, which is overlain by sandy outwash of the older Wellfleet plain and eolian sand (Fig. 9). This section was first described by Oldale et al. (1968) as the Newcomb Hollow clayey silt locality. Oldale et al. (1968, p.4) observed that the "clayey silt, although highly deformed, is texturally representative of the clayey silt found elsewhere in the older Wellfleet plain deposits. It varies laterally from a massive, poorly sorted, till-like deposit to well-laminated clayey silt, silt, and clay. Angular to subrounded pebbles and cobbles are found in both the till-like and well-laminated facies. Boulders are scattered throughout the clayey silt."

In 1984, four sections were measured at this exposure at a scale of 1:16 (Fig. 10). These data plus detailed mapping between sections were used to construct the outcrop map shown in Fig. 9. The sections exposed in 1987
Figure 9. (Stop 3) Lithologic map of Newcomb Hollow Beach north section, prepared in 1985. Inset shows generalized stratigraphy. Sections 1-4 refer to locations of measured stratigraphic profiles shown in Figure 10. In Figure 9, for simplicity all sandy facies are stippled, and large bodies of mud (F-facies) are shown in black. Note diapiric character of loading at contact between diamicton lithofacies and older Wellfleet Plain deposits.
### FACIES CODE

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Figure 10A. (Stop 3) Explanation of facies codes used in Figure 10A (after Eyles et al., 1983).
Figure 10B. (Stop 3) Individual stratigraphic profiles measured at the north Newcomb Hollow Beach outcrop.
differed somewhat due to slumping and exposure of new material. The diamicton facies consists mainly of silty stratified diamicton (Fig. 6) with interbedded sand and mud. Stratification in the diamicton, defined by stringers of micaceous silt and sand, dips to the northwest (Fig. 8B). Massive diamicton is less abundant, and occurs mainly at the base of the section and as isolated clots and lenses within the stratified diamicton. Contacts between the massive and stratified facies are diffuse and conformable. Massive and stratified diamictons are rich in muscovite, which imparts a sparkle to hand samples. Munsell colors of the diamictons are greenish gray (5GY 5/1) to dark greenish gray (5BG 4/1).

The sandy interbeds within the diamicton facies are variable in their thickness and continuity. Ripples and horizontal lamination are common. The upper and lower contacts are typically erosional. Beds of massive (i.e. structureless) sand are also present, which may achieve thicknesses of more than 2 m (Fig. 10, section 4). Soft sediment deformations are common in sandy interbeds, and include convolute bedding and diapiric injections of sand into diamicton (Figs. 11, 12).

Beds of massive and laminated mud occur sporadically throughout the diamicton facies. The mud is identical to the matrix of the massive diamicton in terms of its color, texture and abundance of muscovite. Mud typically occurs interbedded with sand; in profile 4 (Fig. 10), thin (10 cm), horizontally laminated sand and mud layers form a series of rhythmites over a vertical interval of 1.5 m.

The pebble lithologies of the massive and stratified diamicton facies were analysed separately, and no significant differences were observed. For simplicity, the data for both facies have been grouped together for the north
Figure 11. (Stop 3) Diapiric injection of sand into stratified diamicton, photographed in 1987. The sand body trends N10W and dips 70° east. Staff is calibrated in 30 cm divisions.
Figure 12. (Stop 3) Soft-sediment deformations in the diamicton facies. Upper part of photograph shows a ball-and-pillow structure at the contact between sand at the base of older Wellfleet plain deposits and sand at the top of the diamicton lithofacies. The base of the ball-and-pillow structure marks the contact and is highlighted by the dashed line. The solid line below the ball-and-pillow outlines a flame structure of stratified diamicton (overturned to the south) intruded into sand of the diamicton facies. Location is between sections 1 and 3, Figure 9. These structures, photographed in 1986, were not visible in 1987.
and south sections (Fig. 3). The pebble lithology suite of the diamicton facies is notably different with respect to deposits of the older and younger Wellfleet plains, the Eastham plain, and the diamictons at Coast Guard Beach (Fig. 3). In contrast, phyllite predominates in the diamicton facies at Newcomb Hollow Beach. Phyllite fragments are characteristically green, angular and micaceous. Granite and felsic volcanic rock fragments make up the bulk of the remainder of the pebble suite of the diamicton facies.

The granulometric composition of the massive and stratified diamictons is dominated by micaceous silt (Fig. 6; Table 1). With respect to the massive diamicton, stratified diamicton is enriched in sand, and contains about the same amount of clay. The pebble content of the diamicton facies is too low to permit systematic investigations of pebble fabric, which was conducted at only one site, a large lens of pebble-rich massive diamicton (Figs. 8C,D, 9: 15-45 m). This pebble fabric (Fig. 8C,D) shows a scattered but non-random pattern, with most A-axes dipping generally to the west. C-axes are mostly steeply dipping. Striations are common on pebbles.

The contact between the diamicton facies and sandy older Wellfleet plain deposits is strongly deformed by loading. Evidence of loading includes diapiric injections of diamicton, sand and mud into the older Wellfleet plain sandy outwash and ball-and-pillow structures (Figs. 9, 12). Elsewhere the contact is occasionally marked by a boulder lag (Fig. 13). The boulders are typically rounded.

Older Wellfleet plain outwash at the north section consists almost entirely of sandy sediment. A single bed of massive gravel occurs near the top of profile 4 (Fig. 10). Current structures include ripples, trough cross-bedding, horizontal lamination, and planar cross-bedding. Paleocurrent
Figure 13. (Stop 3) Boulder lag concentrate (below dog) at contact between diamicton facies and older Wellfleet Plain outwash, photographed 1987, and located at measured section 3 (Figure 9). Canine is approximately 60 cm high at shoulder.
measurements were not obtained in sandy outwash at the north section. Oldale (1968) reported planar foreset beds dipping west elsewhere in the Wellfleet quadrangle, which is consistent with the paleocurrent data observed at Stop 2 (Fig. 8B).

North of Newcomb Hollow Beach, in the North Truro quadrangle, Oldale et al. (1968) also reported planar cross-bedded sand below outcrops of "clayey silt" believed to be equivalent to clayey silt (i.e. diamicton) at Newcomb Hollow Beach. Moreover, a layer of ventifacts occurred at the contact between the clayey silt and sand. This aspect of Wellfleet plain stratigraphy was not seen at Newcomb Hollow Beach during the summers of 1984-1987. The earlier observations are none-the-less important, for they illustrate another complexity in the behavior of the South Channel lobe during deposition of Wellfleet plain sediments.

INTERPRETATION

Diamicton Facies

Oldale's (1968, 1982) interpretation of the Wellfleet plain as a deltaic body of outwash deposited by the South Channel lobe was based on the geomorphology of the uncollapsed surface of the outwash plain (Fig. 2) and the westerly dip of deltaic foreset bedding in sandy outwash (above). The earlier work of Oldale et al. (1968) did not address the origin of the clayey silt deposits and their relationship to the South Channel lobe. The nature of the Newcomb Hollow Beach outcrops demonstrates that the debris delivered to the terminus of the South Channel lobe was very heterogeneous, and included silty diamicton in addition to sandy outwash. The general style of deposition may have been similar to depositional models for temperate glaciers suggested by
Gustavson and Boothroyd (1987, their Fig. 9E) and McPherson et al. (1987). Their models have been modified to account for the conditions that may have existed at the margin of the South Channel lobe when it abutted glacial Lake Cape Cod in the Wellfleet area (Fig. 16).

The model initially involves the deposition of a subaqueous proglacial sediment apron predominantly consisting of silty diamicton. The abundant striated pebbles found in the diamicton facies suggest that it was derived primarily from basal debris. The material constituting the diamicton facies was then delivered to the margin of the South Channel lobe by subglacial and englacial streams (Gustavson and Boothroyd, 1987). Upon release from the glacier, silty diamicton as well as sorted sand and silt was reworked downslope as repetitive glaciogenic debris flows, which produced stratified diamicton (cf. Evenson et al., 1977). Turbulent eddies produced during flowage probably elutriated silt from the bodies of debris flows, allowing a residual enrichment of sand in stratified diamicton (Table 1; Broster and Hicock, 1985). Additional sand may have been admixed and homogenized into the stratified diamicton by turbulence and shear during flowage of debris as a plastic or viscous-fluid mass (Evenson et al., 1977; Hicock et al., 1981; Broster and Hicock, 1985; Wright and Anderson, 1982). In this model winnowed silt would have been redeposited from suspension as thin beds of massive or stratified mud. The rhythmites occurring in profile 4 (Fig. 10, 7.5-9 m) probably are turbidites consisting of suspension deposits of mud (Bouma interval T_e) interbedded with horizontally laminated sand (T_b or T_d) laid down by tractive currents. The few occurrences of rippled sand (Sr) are also inferred to have originated from turbidity currents (T_c).
The origin of beds of massive sand is more problematical (Fig. 10, profile 4, 1-4 m, profile 1, 7.5-9 m). These layers may reflect deposition by sandy, high-density turbidity currents, in which the sand was deposited from suspension so quickly that neither a bed-load layer nor a traction carpet could form (Lowe, 1982). Alternatively, massive beds of sand may represent suspension deposits from plumes of sandy sediment dispersed in the lake by englacial or supraglacial streams (Sharpe, 1987).

Massive diamicton was probably deposited by coherent flows of debris that did not undergo significant winnowing and that were relatively unaffected by tractive current erosion or deposition. The large body of massive, pebble-rich diamicton between profiles 3 and 4 (Fig. 9) probably originated in this manner. Movement of this diamicton body was sluggish enough that only a weak parallel fabric developed (Fig. 8C,D). Because this particular body of diamicton is deformed by loading (Fig. 9: 15-40 m), it is also possible that the pebble fabric (Fig. 8C, D) has been reoriented. Moreover, at least some massive diamicton at the north section could have been deposited from suspension with sporadic input of ice-rafted debris (cf. Domack and Lawson, 1985).

The soft-sediment deformations that are common throughout the diamicton facies are interpreted to be primarily the result of loading by sandy sediments of the Older Wellfleet plain while the diamicton facies was still in a hydroplastic state. It is likely that at least some of the deformations may also have developed in response to earlier episodes of loading that followed subaqueous debris flows and turbidity currents (e.g. Lowe, 1982; Banerjee, 1973).
Older Wellfleet plain outwash

Older Wellfleet plain outwash at Newcomb Hollow Beach consists mainly of current-bedded sand (Fig. 10: St, Sh, Sp) and local, discontinuous beds of massive gravel (Gcm). Oldale (1968, 1982) interpreted these sediments to be of deltaic origin primarily based on geomorphic considerations (above). The nature and occurrence of sandy bedforms recognized at Newcomb Hollow Beach are typical of delta foreset sequences (Oldale et al., 1968; Gustavson et al., 1975), so only a few additional comments will be offered here. Rippled sand containing lenses of stratified diamicton (Fig. 10, profile 3) reflects reworking of the diamicton facies by meltwater streams. This phenomenon was also seen at Stop 2, and is probably common near the contact between the outwash and diamicton facies.

Subaerial conditions during deposition of the older Wellfleet plain are implied by the ventifacts found occasionally in the outwash, and possibly also by the boulder lag atop the diamicton facies (Fig. 13), which may reflect reworking of the diamicton facies in a beach environment.

STOP 4: Newcomb Hollow Beach (South Section)

The south section occurs 600 m south of profile 4 (Fig. 10), and consists of a small outcrop of massive and stratified diamicton (Fig. 14). Massive diamicton occurs as small exposures at the base of the section, and as a diapiric mass at the north end of the outcrop; the remainder of the outcrop consists of brown (10YR 5/6-6/6) stratified diamicton. Stratification is diapirically deformed around the protuberance of massive diamicton at the north end of the outcrop; at the south end, stratification is also deformed in a convoluted manner. Stratification with flow folds (sense of Evenson et al.,
Figure 14. (Stop 4) Lithologic map of south Newcomb Hollow Beach outcrop, prepared from field sketches and photographs taken in 1987. Explanation as in Figure 10A, except as noted. Inset stereogram shows trend and direction of plunge of axial planes of flow folds.
1977), near the middle of the outcrop (Figs. 14, 15), indicates a southerly paleocurrent. Massive and stratified diamictons at the south section are granulometrically similar to their counterparts at the north section (Fig. 6).

**Interpretation**

The origin of massive and stratified diamicton at the south and north sections is probably similar. The soft sediment deformations at the south section also developed in response to loading by older Wellfleet plain outwash. The most salient difference at the south section is the change in paleocurrent direction, toward the south, whereas at the north section bedding dipped more to the west-northwest (Fig. 8B). This variation at the south section may reflect a local perturbation in flow conditions which was unrelated to westward flowage at the north section. Alternatively, the south section may be an outcrop at southern end of a subaqueous fan or delta, whereas the north section originated nearer the central part of the same body, where flowage was in a more westerly direction.

**Depositional Environment of the Outwash and Diamicton Facies in Wellfleet**

Wellfleet plain outwash is interpreted to have prograded westward as a braid delta upon the diamicton facies (Fig. 16). The soft sediment deformations in the diamicton facies accompanied onloading of the latter by older Wellfleet plain outwash. Braid deltas have been defined by McPherson et al. (1987, p. 331) as "gravel-rich deltas that form where a braided river system progrades into a standing body of water." The braid delta depositing the Wellfleet plain is believed to have been a sandy head-of-outwash (e.g. Evenson and Clinch, 1987) confined to the north and south, respectively, by local re-entrants of the Cape Cod Bay lobe and South Channel lobe (Fig. 1).
Figure 15. (Stop 4) Flow folds in stratified diamicton facies at the south section, photographed in 1987. Staff is calibrated in 30 cm divisions.
Figure 16. Hypothetical model of deposition of diamicton and outwash in the Wellfleet area. (A) Sandy outwash is deposited on a braid delta abutting glacial Lake Cape Cod at the margin of the South Channel lobe. Local winds produce ventifacts on the surface of the outwash plain. (B) Diamicton facies is deposited as subaquatic glaciogenic debris flows in deeper water conditions than (A). Here, mainly silty basal debris is introduced at the glacier margin. The diamicton facies at Coast Guard Beach may have been deposited in a similar manner, but from debris that was richer in sand than diamicton at Newcomb Hollow Beach. (C) Relatively lower lake levels accompany re-establishment of a braid delta, as additional sandy outwash of the older Wellfleet plain is deposited upon the diamicton facies. The latter, still in a hydroplastic state, readily deforms due to loading.
The definition of the braid delta, vis-a-vis glaciofluvial systems, clearly implies the existence of an outwash plain, or sandur, abutting a glacier margin. This requirement is compatible with the geomorphic expression of the Wellfleet plain as a pitted outwash plain (Oldale, 1976). The southwestward gradient of the uncollapsed surface of the Wellfleet plain suggests that its source was some distance offshore from modern Cape Cod (Oldale, 1982). The deltaic component of the braid delta is reflected by deltaic foreset bedding and outwash cones (this study; Oldale, 1968).

In the Wellfleet area, strandlines associated with glacial Lake Cape Cod have not been observed, and hence the nature of lake level fluctuations is unknown. The presence of ventifacts at the top of deltaic sand below the diamicton facies implies subaerial conditions. Later, a subaqueous environment prevailed during deposition of the diamicton facies. The depositional environment was again partly subaerial as the older Wellfleet plain formed.

Comments on the Provenance of the Diamicton and Outwash Facies

The differences between the pebble suites of the outwash and diamicton facies at Coast Guard and Newcomb Hollow Beach are inferred to reflect contrasting provenances of debris delivered to the margin of the South Channel lobe at the time each facies was deposited. The provenance of the South Channel lobe is the Gulf of Maine, whose underlying bedrock is known only in general terms (Ballard and Uchupi, 1975). The pebble suites of the outwash and diamicton facies are representative of the bedrock onshore in southeastern New England (Figs. 1, 3). This phenomenon clearly demonstrates that the South Channel lobe eroded and incorporated similar rocks which must underlie the
Gulf of Maine. Specific indicator pebbles include red sedimentary rocks from Triassic basins, *Obolus* quartzite pebbles from Pennsylvanian conglomerates, and red-to-purple felsites from Siluro-Devonian volcanic rocks (Howe, 1936; cf. Cameron and Naylor, 1976).

Diamicton from the Newcomb Hollow Beach sections is distinguished by a preponderance of phyllite and a lack of granite and quartzite as compared to Wellfleet plain outwash. Diamicton at Coast Guard Beach is enriched in slate, slate-bearing quartz, and quartzite as compared to Eastham plain outwash, which is dominated by felsic volcanic rock and granite. If the diamicton and outwash facies had a common source at each site, it is possible that granite, felsic volcanic rock and quartzite are enriched in the respective outwash facies by virtue of their durability relative to that of phyllite and slate. However, the lithology of ice streams contributing to head-of-outwash ice margins can be extremely variable (Evenson and Clinch, 1987). Hence, it is possible that the lithologic diversity of the pebble suites at each site reflects differences in provenance among the individual ice streams that fed the margin of the South Channel lobe.

The silty matrix of the diamicton facies most likely reflects lake sediments incorporated and reworked by the South Channel lobe. The nature of these materials was probably similar to glaciolacustrine sediments presently found on Cape Cod itself, which consist of silty, micaceous, massive to laminated muds deposited in glacial lake Cape Cod (Sayles and Knox, 1943; Oldale et al., 1967; Oldale, 1982). The ultimate source of the abundant muscovite inherited by the diamictons and muds at Newcomb Hollow and Coast Guard Beaches was probably comminuted phyllite from the basal debris zone of the South Channel lobe.
References


Woodworth, J. B., and Wigglesworth, E., 1934, Geography and geology of the region including Cape Cod, the Elizabeth Islands, Nantucket, Martha's Vineyard, No Man's Land, and Block Island, 322 p.

LATE QUATERNARY GLACIAL AND VEGETATIONAL HISTORY
OF THE WHITE MOUNTAINS, NEW HAMPSHIRE

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INTRODUCTION

The purpose of this field trip is to visit and evaluate critical sites for understanding late Quaternary glaciation, deglaciation, and climate of the White Mountains of New Hampshire. A long-standing controversy concerns the sequence and timing of cirque and continental glaciation in the area, which is important for modeling paleoclimate during the most recent glacial-interglacial transition. Bogs and lakes in the area not only provide meaningful paleoecological records of postglacial climate, but also provide our only significant dating control for deglacial events. Although mountains in eastern United States are too low to provide geomorphic evidence for recent climatic change, tree rings along altitudinal gradients provide excellent records of environmental change during the last few hundred years. Finally, the White Mountains provide an excellent laboratory for monitoring modern-day atmospheric pollution and for evaluating endangered alpine plants and subalpine trees. In the tradition of the American Quaternary Association, we strive to make this field trip an interdisciplinary effort in understanding our environment, past and present.

Our honorary field trip leader is Dr. Richard P. Goldthwait, professor emeritus at Ohio State University, and founder and first director of the University's Institute of Polar Studies (now the Byrd Institute of Polar Studies). His name is synonymous with Quaternary geology in New Hampshire; without his decades of research in and around the White Mountains, this field trip would have little meaning. Dick Goldthwait has led field trips to the Mt. Washington area since the 1940's; in this present effort we have made extensive use of materials from guidebooks for two of his more recent trips: the International Symposium on Antarctic Glaciological Exploration, 1968, and The Friends of the Pleistocene, 1970. More recent geological efforts by numerous workers in the area have merely expanded on his cornerstone work. In 1986, the Quaternary Geology and Geomorphology Division of the Geological Society of America bestowed to Dick Goldthwait their first Distinguished Service Award, an award destined to become the most coveted in the profession.

Other recent field trips to the Mt. Washington area include an excursion of the International Botanical Congress from Montreal, Quebec (Löve and Löve, 1959), and the 40th reunion of the Friends of the Pleistocene, which focused on glacial features in the lowlands just southeast of the White Mountains (Newton, 1977). Some of the stops from those two field trips will be revisited on Days 1 and 3 of the AMQUA trip.

The stops, figures, and tables in this guidebook are numbered sequentially for each day of the trip, in the event that weather conditions make it necessary to reverse the order of Days 1 and 2. The Location Map accompanying this introduction shows the stops, major highways, and other features related to the field trip. The discussion leaders and/or authors for stops described in the guidebook are indicated in parentheses in the heading for each stop.
ACKNOWLEDGMENTS

The authors are especially grateful to Catherine Stultz, Maine Geological Survey, who word-processed the entire text for this guidebook, and to Robert Johnston (also MGS) who assisted in preparing camera-ready figures. The staff of the Mt. Washington Observatory, particularly Guy Gosselin, helped arrange activities on the summit of Mt. Washington, and generously made available the copies of R. P. Goldthwait's bulletin on the geology of the Presidential Range that were distributed to trip participants.
Field trip stops are numbered for each of the three days of the trip. For example, "3-2" indicates Stop 2 on Day 3.
DAY 1

MT. WASHINGTON

Beginning of trip - Glen House, Pinkham Notch, 495 m a.s.l.

Late Wisconsinan continental ice probably covered the highest peaks of the Presidential Range, including Mt. Washington with a summit elevation of 1917 m above sea level (a.s.l.). The cols between peaks have striae and roches moutonnées oriented 140° (Goldthwait, 1940, 1970a,b). In the area around the Glen House, striae are oriented 180-200°, which leads to the question of whether or not these erosional features are contemporaneous (Goldthwait, 1970b).

STOP 1 (R. W. Spear) - Transition from deciduous to coniferous forest, 750 m a.s.l.

Vegetational zonation

Bormann and others (1970) reviewed a number of schemes devised to split the White Mountains vegetation into zones. The only vegetation change recognized by all schemes is the transition from deciduous to coniferous forest at 750 m. Bormann and others (1970) concluded that the major forest trees are individually distributed along the mountain slopes, and they described the upper elevational limit at which each tree species formed an important component of the vegetation.

Zonation schemes of Bormann and Nelson (1963) and M. B. Davis (personal comm., 1980) recognize five zones: (1) northern hardwoods (hemlock phase); (2) northern hardwoods (spruce phase); (3) subalpine (spruce-fir phase); (4) subalpine (fir phase); and (5) alpine. The first zone is called mixed hardwoods by M. B. Davis (personal comm., 1980), and it extends up to about 450 m in elevation. Three species, white pine (Pinus strobus), red oak (Quercus rubra), and hemlock (Tsuga canadensis), reach their maximum in this zone. The second zone, the northern hardwoods (spruce phase), spans the elevations from 450 to 750 m. The altitudinal limits of three species, yellow birch (Betula lutea), sugar maple (Acer saccharum), and beech (Fagus grandifolia), are reached in this zone. The spruce-fir forests of the third zone are found between 750 and 1220 m. Above 1220 m red spruce becomes rare and balsam fir dominates the forest. Fir forest forms the fourth zone between 1220 and 1500 m. Fir and black spruce (Picea mariana) krummholz are found in the lower part of the alpine zone, the fifth zone, but no upright trees occur much beyond 1500 m.

STOP 2 (L. E. Conkey, M. B. Keifer, K. D. Kimball) - Red spruce decline in northeastern United States

The existence, extent, and possible causes of a decline in red spruce in the northeastern United States are being investigated by many researchers. Some areas, such as Camel's Hump in Vermont, appear to have suffered large decreases in spruce basal area, and recent ring widths are quite narrow compared to previous decades (cf. work by Vogelmann, Sicamass, Johnson, etc.). Other locations are not as visually affected, but may show signs of changing environmental relationships and conditions in less obvious ways. Here we
briefly report on work in progress at three locations: Elephant Mtn. (on the Appalachian Trail in western Maine), Mt. Washington, and Mt. Moosilauke (on the western edge of the White Mountains).

At Elephant Mtn., studies of patterns of ring widths and latewood densities of red spruce in an old-growth stand indicate a changing picture of growth, particularly in the wood density record (Figure 1), where the last few decades show decreased amplitude of variation along with lower density values overall (Conkey, in review). After a comparison with a climatic factor that relates well to the overall red spruce record (Conkey, 1986), estimated densities were compared to measured densities, and a running series of 25-year correlation coefficients were calculated and plotted (Figure 2; Conkey, in review). The sharp drop-off in recent decades in an otherwise strong density/climate relationship indicates a marked change in that relationship, and may be related to a complex interaction of climatic events with human-induced environmental changes.

Red spruce decline on Mt. Washington was investigated by surveying the forest by elevation and examining the annual radial growth of red spruce. Tree cores were obtained from forest study plots on Mt. Washington along four transects: a valley and a ridge transect for both the east and west slopes of the mountain. A site consisted of four 0.2 ha study plots. Sites were located at 800, 900, 1000, and 1100 m elevations along each transect, and two sites were located at 1200 m for a total of 18 sites. Cores were taken during 1985 at 1.3 m height from 5 randomly selected red spruce >10 cm dbh at each site (Kimball and Keifer, 1988). These sites represent the elevation range in which red spruce is generally found on Mt. Washington.

The red spruce survey revealed that the basal area of dead red spruce increases with elevation while the number of dead stems increases only to 1100 m (Figure 3). Red spruce in the declining category generally increase with elevation (Kimball and Keifer, unpublished data). In a similar survey of red spruce on Mt. Moosilauke (southern White Mtns.), for live trees only, severely declining trees (>50% needle loss) do not appear to increase with increasing elevation (Figure 4; Conkey and Keifer, unpublished data).

The annual radial growth patterns of spruce are highly statistically correlated \( r > 0.80 \) among the various elevations sampled. Figure 5 shows the mean standardized ring-width index series for 51 years for Mt. Washington. A decline in annual growth starting in 1959 is preceded by a steady increase in growth since 1933 (Kimball and Keifer, 1988). There are various hypotheses to explain the recent red spruce decline which may or may not be applicable to the red spruce on Elephant Mtn., Mt. Washington, and Mt. Moosilauke. Some of these hypotheses include: (1) climatic factors such as drought, wind and winter damage to trees, as well as change in temperature patterns; (2) insect infestation and disease; (3) forest maturation; and (4) atmospheric pollutants, including increased acidity of rain and cloud water, and ozone.

Recent research on subalpine fir forests (Sprugel, 1976; Reiners and Lang, 1979) has greatly increased our understanding of these ecosystems. The subalpine forests are species-poor. Fir makes up 99% of the tree layer; only a few trees of paper birch, red and black spruce, and mountain ash occur in
Figure 1. Red spruce averaged tree-ring widths (upper graph) and densities (lower graph) for Elephant Mtn. In the upper graph, both earlywood width (lower line) and total ring widths are shown. The lower graph shows the complete period mean. From Conkey, 1984, USDA Forest Service Gen. Tech. Rept. NE-90, p. 69-75.

Figure 2. Spearman rank correlation coefficients, calculated on moving 25-year increments, from 1899 to 1984, of measured and estimated maximum densities of red spruce at Elephant Mtn. X-axis shows the central year of each 25-year increment. The estimates derive from a regression from 1899-1945 with April-May temperatures at Farmington, Maine. The horizontal line represents the value of significance at p=0.05, for n=25. From Conkey, in review.
Figure 3. Red spruce condition on Mt. Washington by elevation for basal area (m$^2$/ha) and stems (#stems/ha). For 800-1100 m elevations, n=20. For 1200 m, n=10. From Kimball and Keifer, unpublished data.
Figure 4. Red spruce crown condition on Mt. Moosilauke by elevation and aspect. "% loss" refers to foliage. For each site, n=19-20. From Conkey and Kimball, unpublished data.

Figure 5. Average tree-ring index chronology for red spruce on Mt. Washington for 1933-1984 (n=90). From Kimball and Keifer (1988).
these forests. The diversity of shrubs and herbs is also low, but the forests are rich in bryophytes and epiphytic lichens.

Reiners and Lang (1979) believe that the general patterns of variation they find in subalpine fir forests are related to wind. The patterns associated with increasing altitude are caused by increased wind exposure at higher elevations. Canopy height decreases as elevation increases. The density of fir trees also declines with elevation, as does variation in the age of trees. Reiners and Lang (1979) also identified six patterns of local variation caused by disturbance in the fir forest: waves, strips, hurricane patches, broken tree gaps, glades and avalanche slopes. Sprugel (1976) studied extensively the phenomena of fir waves in the subalpine forests of the northeast. He discovered that trees died at a much faster rate in the exposed front of the fir wave and concluded that this was caused by high wind speed at the front.

STOP 4 (R. W. Spear) - Treeline, 1310–1525 m a.s.l.

Antevs (1932) noted that treeline in the Presidentials depends primarily on altitude; however, it also varies with steepness of slope and exposure to the prevailing wind. He found that treeline on the west slopes of the Presidentials is between 1465 and 1525 m, but it is as low as 1310 m on exposed northwest slopes.

The climatic factors that determine the position of treeline in the White Mountains have been the subject of much research over the years. Climatic factors such as temperature, wind, snow depth, amount of light, and carbon dioxide deficiency, used to explain the position of treeline in other regions, are summarized by Daubenmire (1954), Wardle (1974), and Elliot (1979). In order of importance, Antevs (1932) lists wind, temperature, snow cover, and soil as the main factors influencing treeline in the White Mountains. Wind exposure is considered by many others to be one of the most important factors in determining the position of treeline on the peaks (Monahan, 1931; Griggs, 1942, 1946; Bliss, 1963; Harries, 1966; Tiffney, 1972; and Reiners and Lang, 1979).

STOP 5 (R. W. Spear) - Alpine meadow, 1450 m a.s.l.

The alpine zone lies above 1450 m in elevation. Antevs (1932), Bliss (1963), Harries (1966), and Tiffney (1972) have all made detailed vegetation studies in this zone. They recognized a number of different plant communities distributed along environmental gradients. Bliss (1963) emphasized two environmental gradients in the distribution of alpine plant communities. The first is a gradient of increasing moisture and fog. It is related directly to increasing altitude. Bliss hypothesized that the dominance of Carex bigelowii on the cone of Mt. Washington results from its high photosynthetic efficiency. Bliss' second gradient of snow depth and snow melt is directly related to wind exposure.

Bliss (1963) recognized nine plant communities in the alpine zone: (1) sedge meadow, (2) sedge/dwarf-shrub heath, (3) sedge/rush/dwarf-shrub heath, (4) dwarf-shrub heath/rush, (5) dwarf-shrub heath, (6) Diapensia, (7) snowbank, (8) streamside, and (9) bog. The dwarf-shrub heath is best developed around Lake of the Clouds and at low elevations near treeline. It
is at the low end of Bliss' fog and atmospheric-moisture gradient. **Diapensia** communities are found on peaks and ridges subject to high wind and low snow cover. This community represents one extreme of the snow depth-snowmelt gradient. The other extreme is represented by snowbank communities found in the lee of krummholz and in depressions near treeline.

STOP 6 (R. P. Goldthwait) - Chandler Ridge turnout, Auto Road, at 6.00 miles, 1540 m a.s.l.

Till and scattered erratics at the summit of Mt. Washington and on other summits of the Presidential Range include granite and other igneous rock types (Goldthwait, 1930, 1970a), where local bedrock is schist (Billings and others, 1979), as seen in outcrops along the Auto Road. These scarcely weathered erratics show that the entire range was overridden by the late Wisconsin continental ice sheet. In cirques of the Presidential Range (Figure 6) roches moutonnées, drift of alien provenance, and an absence of looped end moraines have long been taken as evidence that local glaciers did not occupy the cirques after ice-sheet deglaciation (J. W. Goldthwait, 1913, 1916; R. P. Goldthwait, 1940). At the remaining stops today, and at the Durand Lake locality on Day 2, we will examine evidence for the above sequence of glaciation.

To the north one may view the northern peaks of the Presidential Range (from right to left, Mts. Madison, Adams, and Jefferson) and the Great Gulf with its numerous cirques. The northern peaks were not as well rounded by continental ice as were peaks to the south including Mts. Washington and Monroe. Granitic erratics are found on the west slopes but not the east slopes of all of the peaks in the Presidential Range. Goldthwait (1970b) suggested that this "shadow zone" perhaps resulted from dead ice. Davis (1976) made similar observations on Mt. Katahdin in east-central Maine.

The local glacier that occupied the cirques and U-shaped valley of the Great Gulf was probably about 8 km long, suggesting a snowline at about 1000 m (Figure 7), a summer temperature about 8°C cooler than that today (Goldthwait, 1970a, 1970b). Some valleys, such as the Chandler Brook valley below you, however, are V-shaped, suggesting lack of snow and ice accumulation during mountain glaciation or extensive postglacial fluvial erosion.

Although bedrock outcrops along the Auto Road below this point are commonly polished and striated, bedrock between here and the summit of Mt. Washington is more broken by frost action and slopes are more covered with felsenmere, including block stripes, polygons, and lobate terraces.

STOP 7 (R. P. Goldthwait) - "Cow Pasture", Auto Road, at 6.75 miles, 1740 m a.s.l.

Block polygons (Figure 8) and stripes (Figure 9) are well exposed here, as are tundra soils. Goldthwait (1968) suggested that the occurrence of sorted patterned ground, such as fissure and polygon nets, stretched polygons, block stripes, striped terraces, and lobate terraces, are a function of steepness of slope and content of fines (silt and clay) in till. Thus, the above forms are listed in order of increasing steepness of slope and/or increasing wetness of substrate, which is a function of the percentage of fines in till (diagram will be provided separate from guidebook). From soil movement studies on Mt.
Figure 6. Surface features of the Presidential Range (from R. P. Goldthwait, 1970a).
Figure 7. Profile of Mt. Washington and northern Presidential Range from south on the right (Oakes and Dry Gulfs) to north on the left (Bumpus Basin), showing present annual and mean monthly summer temperatures at different elevations on the left. Former temperatures are shown on the right. The vertical relief is greatly exaggerated, and all peaks have been consolidated diagrammatically into one.
Figure 8. Block nets on Bigelow Lawn, Mt. Washington (from R. P. Goldthwait, 1940).
Figure 9. Block stripes on Bigelow Lawn, Mt. Washington (from Goldthwait, 1940).
Washington, Goldthwait (1968) noted that blocks in large patterned ground features, such as stripes and polygons, were essentially stable, based on less than 2 cm movement in 32 years and a mature lichen cover. However, Goldthwait (1968) also found that miniature patterned-soil forms on flat areas showed movement up to 6 cm in a year.

Despite intense cryoturbation of the till in tundra areas of the Presidential Range, mature pedogenic profiles have developed (George D. Bailey, in Goldthwait, 1968). However, these soils generally do not include textural B horizons, although cambic B horizons are present. On Mt. Katahdin's Table Land this lack of clay buildup suggests that the soils could very well have developed during postglacial time, supporting the notion that all upland areas were probably scraped clean by an overriding continental ice sheet (Davis, 1976, 1978, unpublished data). Bailey (in Goldthwait, 1968) tentatively classified the tundra soils on Mt. Washington as typic cryorthods in the U.S.D.A. system.

The irregularity of treeline and krummholz forest in the Presidential Range is exhibited by the bare crest of Chandler Ridge as low as about 1250 m, in contrast to some dwarf spruce reaching as high as 1340 m on the Alpine Garden to the south (we will walk to an overlook of this area). This irregularity suggests that wind and snowbanks (i.e. moisture) are important factors that control treeline in this area.

Above treeline, vegetation primarily consists of alpine heath communities, which reach their peak bloom in late June on the Alpine Garden. Sedges, rushes, mosses, and fruticose lichens are common on stable centers of patterned ground. Several endemic species, including the rare and endangered Potentilla robbinsiana (dwarf cinquefoil) suggested to some early workers, primarily botanists, that the tundra zone of New England mountains must have been refuge areas, hence nunataks, throughout Wisconsin time. However, geological evidence generally refutes this so-called nunatak hypothesis, both in the Presidential Range (Goldthwait, 1940, 1970a) and on Mt. Katahdin (Davis, 1976, 1978, in press).

Excellent publications on alpine flowers of the Presidential Range are for sale at the Appalachian Mountain Club's Pinkham Notch Camp.

STOP 8 (R. P. Goldthwait) - Parking lot at site of former Air Force building, Auto Road, at 7.50 miles, 1845 m a.s.l.

Till, with numerous faceted and striated erratic clasts, was exposed in pits dug for fuel tanks and for construction of a large building for cold-weather experiments by the U.S. Air Force on this site. Permafrost is also found at depth in pits from here to the Summit (Goldthwait, 1968). Patterned ground features similar to those described at Stop 7 are common. This site also offers a fine view of the Great Gulf and the northern peaks of the Presidential Range. From here we will proceed on foot to the summit, paying attention to the relationship between frost-riven block forms and slope angle.
STOP 9 (R. P. Goldthwait, B. K. Fowler, K. D. Kimball, P. T. Davis) - Summit of Mt. Washington, 1917 m a.s.l. (highest point in northeastern U.S.A.)

Goldthwait (1968) described three separate erosion surfaces that may be viewed from the Summit on a clear day. The Presidential Upland is preserved only as an undulating shelf flanking Mts. Washington (Alpine Garden and Bigelow Lawn) and Jefferson (Monticello Lawn) at about 1525 to 1675 m. It could be either a remnant of a Tertiary (?) erosion surface accordant with the tops of lower nearby peaks or, less likely, an altiplanation terrace suggesting rapid erosion during Pleistocene time. Below this surface at the ends of ridges, especially on the west side of the Presidential peaks, are broad valley slopes, perhaps of Pliocene (?) age. Pleistocene V-shaped valleys cut by streams and U-shaped valleys carved by local glaciers are preserved on the southwest sides and the south, east, and north sides of the peaks, respectively. During postglacial time the walls of these valleys have been undergoing mass wasting that includes some large landslides.

Till, with numerous faceted and striated erratic clasts, was also found in pits dug for construction of buildings in the Summit area over the last 100 years. The limited weathering of the erratic clasts (rinds generally less than 2 mm thick), and the lack of weathered debris surrounding these clasts, suggest that continental ice covered Mt. Washington during late Wisconsin time (Goldthwait, 1940, 1968, 1970a, 1970b).

A minimum length of time since Wisconsin deglaciation is provided by radiocarbon ages and pollen assemblages derived from bog- and pond-bottom sediments. The oldest radiocarbon ages from the White Mountains are summarized in Table 1. Based on field work near modern glaciers in southeastern Alaska, Goldthwait and Mickelson (1982) constructed a model for the deglaciation of the White Mountains which involves downwasting of the continental ice sheet, causing the peaks to become ice-free before the valleys. Thus, the oldest radiocarbon age available from Mirror Lake, located on a valley floor in the western White Mountains (Table 1; Davis and others, 1980), suggests that the highest peaks in the Presidential Range probably were deglaciated by 14 to 15 ka, or perhaps even earlier. Although a much younger time of deglaciation might be suggested by a radiocarbon age (Table 1) from Lakes of the Clouds on Mt. Washington (Stop 10), pollen assemblages from below the dated horizon suggest an older age (Davis and others, 1980; Spear, unpublished data).

Black and white photographs taken in 1869 and a mature lichen cover suggest that large, frost-riven blocks in the Summit area have not moved over the last century, except where disturbed by construction, much of which has occurred since 1980. A map of cultural features in 1968 for comparison with the Summit area today will be provided separate from the guidehook.

The Mt. Washington Weather Observatory is the oldest, continuously operated, high-altitude station in North America. Climatic variations over the last several decades are summarized by Kimball and DiMauro (1986).

Time will be available for lunch and browsing on the Summit. A fine museum will be open in the Observatory Building. During the afternoon, a more energetic group will proceed on foot down to Lakes of the Clouds (Stop 10), Tuckerman Ravine (Stops 11 and 12), and on to Pinkham Notch for supper (a
Table 1. Minimum-limiting radiocarbon ages for deglaciation of the White Mountains, from basal organic sediments in bogs and ponds

<table>
<thead>
<tr>
<th>Radiocarbon Age (yr B.P.)</th>
<th>Laboratory Number</th>
<th>Location 1</th>
<th>Elevation</th>
</tr>
</thead>
<tbody>
<tr>
<td>13,870 ± 560&lt;sup&gt;2&lt;/sup&gt;</td>
<td>BX-5429</td>
<td>Mirror Lake, West Thornton</td>
<td>200 m</td>
</tr>
<tr>
<td>13,000 ± 400&lt;sup&gt;2&lt;/sup&gt;</td>
<td>QL-1133</td>
<td>Deer Lake Bog, Mt. Moosilauke</td>
<td>1300 m</td>
</tr>
<tr>
<td>12,870 ± 370&lt;sup&gt;2&lt;/sup&gt;</td>
<td>QL-985</td>
<td>Lost Pond, Pinkham Notch</td>
<td>650 m</td>
</tr>
<tr>
<td>11,530 ± 420&lt;sup&gt;3&lt;/sup&gt;</td>
<td>I-10684</td>
<td>Lakes of the Clouds, Mt. Washington</td>
<td>1540 m</td>
</tr>
</tbody>
</table>

<sup>1</sup>See Location Map in Introduction

<sup>2</sup>From M. B. Davis and others (1980)

<sup>3</sup>From R. Spear (unpublished data)
demanding 8-km walk with about a 1200-m drop in elevation; rough on the knees!). The less masochistic group may remain on the Summit, where a tour of the Weather Observatory will be arranged. Also, this group will have more time to examine patterned ground and botanical features around the Summit, before returning by vehicle to Pinkham Notch.

STOP 10 (B. K. Fowler, P. T. Davis, K. D. Kimball, R. W. Spear) – Lakes of the Clouds, 1555 m a.s.l.

Between the Summit and the Washington - Monroe col numerous short stops may be made to observe patterned ground features similar to those described on earlier stops. Sorted solifluction terraces, or block lobes, are common, and apparently have not moved over the past century (Goldthwait, 1970b). In general, block forms in this area are tear-drop shaped lobes on slopes greater than 11°, crescent-shaped terraces between 7° and 10°, stripes between 4° and 7°, and polygonal nets on slopes less than 3° (Goldthwait, 1970b).

In the area of the Washington - Monroe col Fowler (1971) described three types of evidence for determining ice flow direction. Striated bedrock surfaces and roches moutonnées, especially where oriented oblique to the direction of foliation in the Littleton Formation, suggest that the regional flow of continental ice was from the northwest. However, this general trend may vary by as much as 30° depending on irregularities in the topography (Fowler, 1971). Mt. Monroe, with its gentle side facing northwest and its steep craggy side facing southeast, could also owe its distinctive shape to erosion by continental ice. However, the shape of Mt. Monroe's summit could also, in part, be due to the regional structure of the Littleton Formation, which dips to the northwest in this area (Fowler, 1971).

Two major funded research projects of the Appalachian Mountain Club are ongoing in the Washington - Monroe col. An effort to protect one of the world's rarest alpine plants, Potentilla robbinsiana (dwarf cinquefoil), was initiated when the plant became listed as a Federally Endangered Species in 1981 (Kimball, 1987b). The dwarf cinquefoil is relatively abundant in the Washington - Monroe col because of the high winds, little snowpack, and active soils which inhibit growth by competing plants. Although dwarf cinquefoil is better adapted to these conditions by growing close to the ground and by anchoring itself with a deep tap root, frost action remains an important cause of mortality for the plant (Kimball, 1987b). A long-term monitoring program of atmospheric pollutants to better understand their impact on high-elevation ecosystems in the Northeast also began in 1981 (Kimball, 1987a). This study was unique in that it was the first to properly collect and analyze cloud-water chemistry. Data collected between 1983 and 1986 show both rain and cloud atmospheric events to be acidic, with sulfuric and nitric acids being the primary constituents. Along with other factors, such as climatic change, acid rain and clouds may be a cause for the demise of certain tree species in alpine areas of New England (Kimball, 1987a; Conkey, 1986).

Two lake basins lie 1.7 km southwest of the summit of Mt. Washington. Sediment cores were taken from the larger of the two basins, the "Upper" lake. The lakes are in the alpine zone 75 m above treeline. Patches of fir and black spruce krummholz grow around their shores. Dwarf-shrub heath with Vaccinium uliginosum, Vaccinium vitis-idaea, Ledum groenlandicum, Kalmia polifolia, and dwarf birch (Betula glandulosa) and snowbank vegetation with
Figure 10a. Pollen percentage diagram for Lakes of the Clouds (modified from Spear, 1981).
Figure 10b. Pollen percentage diagram for Lakes of the Clouds (modified from Spear, 1981).
Figure 10c. Pollen percentage diagram for Lakes of the Clouds (modified from Spear, 1981).
Houstonia caerulea var. Faxonorum and Veratum viride are also common around the lake (Bliss, 1963).

The bedrock underlying Lake of the Clouds is interbedded quartzite and mica schist of the Littleton Formation (Billings and others, 1946). The 15-ha watershed of the lake is 30 times the size of the lake. It is long and narrow, extending northeast from the lake to an elevation of 1676 m, and contains several springs (Buchanan, 1975).

The pattern of sedimentation at Lake of the Clouds is complex for several reasons. The watershed is poorly vegetated and steep. Large amounts of material must wash into the lake during storms. The bottom of the lake is also very irregular. Large boulders and crevices complicate deposition. Finally the lake is shallow and easily mixed by high winds. The sediments are concentrated in the center one-half to one-third of the lake. Despite the irregularities in sedimentation, the macrofossil record at Lake of the Clouds does show approximately the same pattern in cores A and B. It is possible to account for the spatial distribution of macrofossils within Lake of the Clouds at any one time; however, it can not be easily determined whether the lake's effectiveness in trapping and preserving macrofossils has remained constant through time. Are there few macrofossils in the upper 5000 years of sediment because the vegetation, particularly krummholz, was less abundant or because the lake became shallower?

Because of upslope transport of pollen, the complete pollen diagram from Lake of the Clouds (Figure 10) provides a good summary of changes in the regional vegetation. Several important stratigraphic features are: (1) the rise in spruce pollen percentages at 11,500 yr B.P., their decline around 10,000 years ago, and a second rise in the last 2000 years; (2) the sharp peak (28%) in alder pollen 10,500 years ago; (3) the rise in birch pollen percentages immediately after the alder peak; (4) the rise of white pine (Pinus haploxylon type) around 9500 years ago; (5) the increasing percentages of hemlock pollen 8500 years ago, their abrupt decline 4800 years ago, and subsequent slow recovery; and (6) the increase in beech pollen 6500 years ago.

The arrival of spruce and fir krummholz at 10,300 yr B.P. marks the beginning of the Holocene. Fir krummholz was probably more extensive at Lake of the Clouds during the first half of the Holocene than it is now. Besides abundant fir and spruce needles, the maximum numbers of Ericaceae, Cyperaceae, and Potentilla tridentata seeds are found in this interval. The seeds indicate that open patches still existed, but extensive krummholz that traps snow (Tifflney, 1972) and creates favorable microclimates for other species had to have grown around the site. Treeline may have been very close to Lake of the Clouds. The climate at Lake of the Clouds may have been similar to that of Eagle Lake Bog at 1280 m or the Horn at 1200 m. If so, the mean annual temperature was 2°C higher.

STOP 11 (P. T. Davis, B. K. Fowler) - Cirque headwall, Tuckerman Ravine, 1550 m a.s.l.

Tuckerman Ravine is known for its late-season skiing, which is possible largely because of the wind-drifted snow from the extensive Bigelow Lawn to the west. A small patch of snow at the foot of the headwall was still skiable on 1 August, 1969, and the entire headwall is commonly skied on Memorial Day.
(late May). In many years during early June the remaining snow at the foot of
the headwall becomes undermined by meltwater from above and a snow arch forms
just south of the Tuckerman Ravine Trail. In the 1920s large patches of snow
commonly survived the entire summer; however, evidence is lacking for small
cirque glaciers reforming during the Little Ice Age of the last few centuries.

R. P. Goldthwait (1940, 1970a) mapped striations and roches moutonnées on
numerous cirque headwalls in the Presidential Range, including some
particularly long, deep grooves on the headwall of Tuckerman Ravine. Thompson
(1960) and Bradley (1981) interpreted these grooves to be the result of snow
and debris avalanches. However, Goldthwait (1940, 1970a) interpreted these
features as evidence for erosion by a continental ice sheet postdating cirque
glaciers, as modern cirque glaciers pluck rather than abrade their headwalls.

STOP 12 (P. T. Davis, B. K. Fowler) - Cirque floor, Tuckerman Ravine, ~1220 m
a.s.l.

In Tuckerman Ravine huge protalus ramparts of locally derived angular
blocks of schist up to 30 m across cover the north side of the upper cirque,
and could obscure end moraines if alpine glaciers were active following ice-
sheet melting, as suggested by Thompson (1960). However, the lower part of
Tuckerman Ravine cirque is not obscured by mass wastage deposits, and does not
exhibit any alpine moraines. A small tarn, Hermit Lake, is ponded by bedrock,
not an end moraine.

The head of Huntington Ravine, the next cirque north of Tuckerman Ravine,
is covered by a huge talus fan of locally derived blocks of schist up to 15 m
across. A few angular schist blocks up to 40 m in length lie on the cirque
to floor beyond the fan and other slope deposits occur along cirque sidewalls.
However, these blocks do not obscure the major part of the cirque floor, thus
do not explain the absence of alpine moraines.

As long advocated by J. W. Goldthwait and R. P. Goldthwait, the abundance
of erratic stones on cirque floors and the absence of distinct alpine moraines
indicate that the most recent substantial mountain glaciation preceded the
last ice sheet occupation. Glacial drift and coarse locally derived clasts,
moved downslope by postglacial debris flows, debris avalanches, landslides,
and streams, is not evidence for mountain glaciers postdating melting of the
last continental ice sheet that covered Mt. Washington.

STOP 13 (R. W. Spear, P. T. Davis, B. K. Fowler) - Lost Pond, Pinkham Notch,
650 m a.s.l.

A radiocarbon age of 12,870 ± 370 yr B.P. (QL-985) from basal organic
sediments in a core recovered from Lost Pond (Davis and others, 1980) provides
a minimum-limiting age for deglaciation of a major glacial through valley in
the White Mountains. For a variety of sedimentological and geomorphological
reasons (Davis and Davis, 1980), this radiocarbon age could also be too young
by hundreds to perhaps even thousands of years. Based on similarities in
pollen assemblages at this site and at other high-elevation sites, including
the Lakes of the Clouds, this radiocarbon age suggests that the White
Mountains underwent rapid deglaciation (Davis and other, 1980), a view that
supports the model of deglaciation developed by Goldthwait and Mickelson
(1982).
Stages in Vegetation Development

>13,000-11,750 yr B.P.

The first period, from before 13,000 to 11,750 yr B.P., was characterized by sparse vegetation and periglacial conditions. The summits and high slopes were practically devoid of vegetation and had a severe climate. Lower in the valleys open "tundra"-like vegetation was prevalent. The sediments at three mountain sites, Lake of the Clouds, Deer Lake Bog on Mt. Moosilauke, and Lost Pond in Pinkham Notch, are highly inorganic, with less than 5% organic matter (Figure 11). Pollen influx is low in sediments from this period, even at Lost Pond (Figure 12). In addition, the two high-elevation sites have low pollen percentages of the poorly dispersed "tundra" taxa (Figure 13).

11,700-10,300 yr B.P.

"Tundra"-like vegetation grew at high elevations during the second period, from 11,750 to 10,300 yr B.P. By this time poplar (Populus) and spruce had invaded the valleys (Figure 14), and extensive woodland had developed on the lower slopes. Sediments at all three sites became increasingly organic during this interval, indicating the establishment of more continuous vegetation. At Lake of the Clouds and Deer Lake Bog, where alpine tundra grew, the percent organic matter increased from roughly 2 to 15%, while at Lost Pond in spruce woodland it rose from 5 to 25%. Pollen influx also increased by 5 to 6 times at Lake of the Clouds, 4 times at Deer Lake Bog, and 18 times at Lost Pond. The percentages of "tundra" taxa reach their highest values at Lake of the Clouds and Deer Lake Bog. At Lost Pond percentages of "tundra" taxa declined steeply at 11,500 yr B.P. as spruce invaded the area and spruce pollen increased. Around 11,000 yr B.P., Juniperus or Thuja reached high elevations and grew as a low prostrate shrub.

10,300-9000 yr B.P.

The third period, beginning with the arrival of spruce at high elevations, dates from 10,300 to 9000 yr B.P. Spruce probably grew in patches of krummholz rather than as forest trees at high elevations. Percentages of "tundra" taxa remain high indicating the open nature of the vegetation during this time. Black spruce was most likely the krummholz-forming species, for it is the species found at treeline today. Around 9600 yr B.P. spruce populations apparently died back. The decline is more evident at Deer Lake Bog. Fir began to replace spruce krummholz between 10,000 and 9000 yr B.P., but it did not become very abundant until after 8500 yr B.P. Although birch reached the high-elevation sites at about the same time as fir, it was never very abundant. Despite high pollen percentages in this period, alder was not an important species in the watersheds of either Lake of the Clouds or Deer Lake Bog. Alder was, however, present at Lost Pond.

9000 yr B.P.-Present

The subalpine fir forests became established around Deer Lake Bog about 9000 years ago. Alpine tundra has always grown around Lake of the Clouds, although fir macrofossils show that krummholz may have been much more
Figure 11. Percentage of organic matter in sediment cores taken from Lakes of the Clouds (Mt. Washington), Deer Lake Bog (Mt. Moosilauke), and Lost Pond (Pinkham Notch). Modified from Spear (1981). Time scale based on radiocarbon ages from Table 1.
Figure 12. Total pollen influx for sediment cores taken from Lakes of the Clouds, Deer Lake Bog, and Lost Pond (modified from Spear, 1981).
Figure 13. Percentage of "tundra" pollen taxa in sediment cores taken from Lakes of the Clouds, Deer Lake Bog, and Lost Pond (modified from Spear, 1981).
Figure 14. Percentage of Picea (spruce) pollen in sediment cores taken from Lakes of the Clouds, Deer Lake Bog, and Lost Pond (modified from Spear, 1981).
extensive prior to 7000 yr B.P. Lake of the Clouds sediments remained fairly inorganic. Loss-on-ignition is approximately 25% throughout the Holocene, indicating lower productivity in its watershed and greater inputs of inorganic sediments than at the other sites. The percentages of "tundra" taxa remained high at Lake of the Clouds and doubled during the last 5000 years. The fossil record shows that the vegetation around Deer Lake Bog, as well as Kinsman Pond and Eagle Lake Bog in the Franconia Notch area, has changed little during the last 9000 years. Spruce has not increased in abundance near these sites during the last several thousand years as it has at lower elevations.

SUMMARY OF VEGETATIONAL CHANGES IN THE WHITE MOUNTAINS, NEW HAMPSHIRE

(1) Analyses of modern surface samples, sediments from several lake basins and from multiple cores within a single basin, demonstrate the difficulties encountered in interpreting the paleoecological record of high-elevation sites. Studies of the modern pollen rain illustrate the massive upslope transport of lowland pollen types and show that pollen from five taxa, Ericaceae, Cyperaceae, Caryophyllaceae, Lycopodium selago, and Houstonia caerulea var. faxonorum, are restricted to alpine pollen assemblages. Recent macrofossil assemblages show that spruce abundance in the modern vegetation is over-represented by the number of spruce fragments found in surface sediments. Comparisons of the two cores analyzed at Lake of the Clouds and of the cores taken at the other sites help us to understand the effects of sediment focusing on pollen influx, and the distribution of plant fragments in small high-altitude lakes.

(2) The White Mountain region was deglaciated prior to 13,000 yr B.P. Downwasting of the continental ice sheet was rapid. The summits projected above the ice sheet as nunataks for only a brief period of time. Residual ice may have existed in Franconia Notch until 11,000 yr B.P.

(3) The record of high-elevation vegetation from late-glacial time and the beginning of the Holocene provides a record of the major climatic changes from 13,000 to 9,000 yr B.P. Once subalpine forests and alpine meadows were established, the two most important climatic factors influencing high-elevation vegetation were wind and moisture. The climatic fluctuations, especially those of temperature, have not been large enough to significantly affect the vegetation during the Holocene.

(a) 13,000 to 11,750 yr B.P.

From over 13,000 to 11,750 yr B.P. a barren periglacial desert covered the highest altitudes in the White Mountains. Tundra vegetation occupied the lower slopes and valleys.

(b) 11,750 to 10,300 yr B.P.

During this period tundra vegetation surrounded all high-elevation sites. The tundra was sparse and consisted of several taxa, particularly Artemisia and Caryophyllaceae, which indicate disturbance. The summits were subjected to intense periglacial activity. At the end of this period shrubs such as Salix, Juniperus, and dwarf birch invaded the tundra at Lake of the
Clouds. At lower altitudes on the slopes and in the valleys, spruce woodland dominated the landscape.

(c) 10,300 to 9000 yr B.P.

At 10,300 yr B.P. black spruce krummholz arrived at Lake of the Clouds and Deer Lake Bog. Macrofossils of fir, birch, and shrubs also occur in sediments of this age. The temperature increased to modern levels or even above. The arrival of krummholz species was delayed about 500 years until 9750 yr B.P. at the Franconia Notch sites. At these sites the establishment of subalpine forests spanned a much shorter period of time. Forests with poplar, spruce, and birch replaced the spruce woodlands of low elevations.

(d) 9000 yr B.P. to Present

Subalpine fir forests became well established 9000 yr B.P. Subsequent changes in the subalpine fir forest, treeline, and the alpine meadow are difficult to determine. There is some evidence at Lake of the Clouds that the fir krummholz was more extensive and treeline higher during the first 4000 years of the Holocene. After 5000 yr B.P. the pollen percentages of alpine indicators increase and the number of fir macrofossils drop. Low-elevation forests give a better record of Holocene climatic fluctuation. Northern hardwood taxa arrived in the first half of the Holocene, but it was not until the expansion of spruce populations approximately 2000 yr B.P. that the modern vegetation zonation became established.
DAY 2

GLACIAL GEOLOGY OF THE NORTHEASTERN WHITE MOUNTAINS
AND UPPER ANDROSCOGGIN RIVER VALLEY

STOP 1 (W. B. Thompson) – Androscoggin Moraine, Shelburne, New Hampshire and
Gilead, Maine

The Androscoggin Moraine is a cluster of end moraines located in the
Androscoggin River valley, on the border between Shelburne, New Hampshire, and
Gilead, Maine (Figure 1). Parts of this moraine system were described by
Stone (1880), and the segment seen at Stop 1 was illustrated in Stone (1899).
Leavitt and Perkins (1935) disagreed with Stone's identification of this ridge
as a moraine, but they failed to present convincing evidence to the contrary.
Thompson (1983) agreed with Stone's interpretation. He named the Androscoggin
Moraine and found additional segments of the moraine that had not been
previously reported.

Detailed mapping of the densely wooded terrain in this area has shown
that the Androscoggin Moraine system includes at least 21 individual moraine
ridges (Figure 2; Thompson and Fowler, in press). Some of these ridges
probably were connected with one another when they were deposited, but have
since been eroded by the Androscoggin River and its tributaries. The
erosional gaps make it difficult to correlate moraine segments across the
valley, but the map pattern of the moraines suggests that they represent
several closely spaced ice-margin positions. A single glacial lobe occupying
the full width of the valley may have divided into three narrower ice tongues
separated by Hark Hill and Crows Nest as the ice receded. Striations and
roches moutonnees provide evidence of topographically controlled east­
southeastward glacial flow in this section of the Androscoggin Valley,
corresponding to the orientation of the moraine system.

The moraine crests rise in elevation from 720 ft (219 m) near the river
to 1260 ft (384 m) where the highest segments terminate against Stock Farm
Mtn. and the hillside east of Mt. Cabot (Figures 2, 3). Some of the moraine
ridges are very steep-sided and up to 30 m high. Numerous large boulders (1-8
m) are strewn along the ridges. Most of these boulders are local rock types
that outcrop just upvalley. Exposures of the Androscoggin Moraine are limited
to a few shallow cuts along logging roads, so backhoe pits were dug in four of
the ridges to determine their composition. The materials found in the test
pits chiefly consisted of sandy, stony diamictons interpreted as flowtills
deposited in an ice-marginal environment. The tills contained deformed lenses
and interbeds of silt, sand, and gravel. Stone counts revealed a predominance
of local bedrock lithologies among the till clasts, but also suggested
differences in provenance on opposite sides of the valley (Thompson and
Fowler, in press).

The Androscoggin Moraine is significant because it is the most clearly
defined of the few end moraines that have been discovered in the White
Mountains, and it is the only known example of a cross-valley moraine. It
indicates that vigorous glacial ice flow persisted in the upper Androscoggin
Valley when the adjacent Presidential, Carter, and Mahoosuc Ranges had largely
emerged from the late Wisconsinan ice sheet. The moraine formed as ice
flowing past Gorham was funneled into the narrow section of the Androscoggin
Figure 1. 1:250,000-scale map showing geographic features and locations of end moraines in the northeastern White Mountains. AM: Androscoggin Moraine, SM: Success Moraine, CM: Copperville Moraine. Numbers indicate locations of stops on Day 2.
Figure 2. Southeastern portion of the Shelburne 7.5-minute quadrangle, showing location of Stop 1 and segments of the Androscoggin Moraine system (heavy black lines). Modified from Thompson and Fowler (in press).
Figure 3. View southwestward across the Androscoggin River valley, showing moraine ridge projecting eastward from Stock Farm Mtn. (center-right to center). River is concealed behind tree-covered moraines in foreground. From Thompson and Fowler (in press).
Valley between the Carter and Mahoosuc Ranges. It is slightly older than the Success Moraine on the northwest flank of the Mahoosuc Range (Figure 1), though both moraines probably were deposited by the same late-glacial ice lobe. There is no evidence that this ice flowed directly down the valley from the lake basins at the head of the Androscoggin River. Instead it is believed to have flowed from the west and northwest as part of an ice lobe spilling out of the upper Connecticut River basin. Striation trends support this source; and Gerath (1978) noted that meltwater from the Connecticut River basin continued to flow eastward (through the Randolph valley) into the Gorham area until the latter was almost totally ice-free.

The age of the Androscoggin Moraine is uncertain, but it is believed to have been deposited close to 14,000 yr B.P. (Thompson and Fowler, in press). This conclusion is based on the ice-retreat models proposed by Davis and Jacobson (1985) and Dyke and Prest (1987). Paleogeographic reconstructions by these authors show the Laurentide Ice Sheet separating over the White Mountains and Mahoosuc Range by 14,000 yr B.P. and retreating to the north side of the Boundary Mountains (on the New Hampshire-Quebec border) by 13,000 yr B.P. According to this model, the Androscoggin Moraine and other ice-marginal deposits in the Berlin-Gorham area would have been formed at the edge of the Laurentide Ice Sheet soon after the adjacent mountains were deglaciated. On the other hand, radiocarbon ages indicate that the ice margin stood at Kennebunk in southwestern coastal Maine as recently as 13,800 to 13,200 yr B.P. (Smith, 1985). If the ice then withdrew progressively northwest from the Maine coast to the White Mountains (as suggested by preliminary mapping results), the upper Androscoggin Valley could have been deglaciated between 13,000 and 12,000 yr B.P. Drawdown of the ice sheet and marine transgression were occurring north of New Hampshire in the St. Lawrence Lowland by this time (Dyke and Prest, 1987), so the Androscoggin Moraine may have been deposited by the ice mass that was left stranded over southeastern Quebec and northernmost New England. This younger age for the Androscoggin Moraine seems less probable because it ignores the separation of the ice sheet over the White Mountains that probably occurred by 14,000 yr B.P. as discussed above. However, additional field work and radiocarbon ages are needed to reconstruct the deglaciation history of northern New Hampshire.

STOP 2 (R. F. Gerath) - Bear Spring Brook Meltwater Channel, Gorham

This significant glacial feature was recently noted by Gerath in a study of aerial photographs. The following discussion offers a number of preliminary conclusions relating to the meltwater channel seen at Stop 2. Bear Spring Brook drains southeastward from the saddle between Pine Mountain and Mount Madison (Figure 4). The brook is underfit and occupies an abandoned meltwater channel. The lowest reach of the 2-km long channel cuts eastward through thick drift, and it is evident that meltwater in this channel ultimately flowed northward down the Peabody River valley.

The Bear Spring Brook channel clearly indicates that meltwater from a body of ice in the Moose River valley flowed into the Peabody Valley. This ice was at least 140 m thick and was almost certainly connected to thicker regional ice issuing from the Connecticut River basin. The continuity of the meltwater channel and its deep incision on the valley floor indicate that it was eroded in a subaerial environment. We will discuss the possibility that part of the channel was eroded subglacially, and that its lower course was
Figure 4. Map showing the Bear Spring Brook meltwater channel and the channel northeast of Pine Mtn. (Carter Dome and Berlin 7.5-minute quadrangles).
diverted by ice occupying the upper Peabody Valley. We will also observe part of a small fluvial-lacustrine ice-contact sequence that was deposited in the Peabody Valley under the influence of the Moose Valley ice. Other bodies of stratified drift and glacial features in this area have yet to be documented by field observations.

A second, less clearly defined, meltwater channel is located on the north side of Pine Mountain (Figure 4). Its threshold is slightly lower (8 m) than the head of the Bear Spring Brook channel. The shallow depth suggests that it was only briefly occupied by meltwater that soon found a lower course and eventually eroded the lower Moose River valley west of Gorham (Gerath, 1978).

The geomorphology of the Bear Spring Brook channel indicates that the lower Peabody Valley was substantially deglaciated while ice north of the Presidential Range was still connected to regional sources. This is consistent with Gerath's (1978) conclusion that the Androscoggin Valley was nearly ice-free while a source of meltwater persisted in the upper Moose Valley.

The outlets of the Pine Mountain drainageways are separated by the drift plug that may have dammed part of the complex sequence of lake sediments in the lower Peabody Valley. The work of Haselton and Fowler (1988) will help clarify the deglacial events in this area. (The Peabody River section will be examined at Stop 1 on Day 3.)

STOP 3 (W. B. Thompson, R. F. Gerath, R. P. Goldthwait) - Summit of Pine Mountain, Gorham

The trail on Pine Mountain (Figure 4) provides excellent vantage points from which to view and discuss the glacial landscapes of the northeastern White Mountains. The degree to which the late Wisconsinan ice sheet remained active during deglaciation of this region continues to be a topic for lively debate. An issue that has developed in recent years is the relative importance of Laurentide versus Appalachian ice in the deglacial history of the area. To answer this question, we require more information concerning the timing of drawdown of the Laurentide ice sheet and marine transgression in the St. Lawrence Lowland, the evolution of late-glacial ice-surface profiles over northern New Hampshire, and radiocarbon ages that limit the time of ice retreat.

Glacial grooves on the summit ledges of Pine Mountain trend 135-142°, parallel to the regional flow of late Wisconsinan ice in this part of the White Mountains. To the south is the Peabody River valley, which is situated between the Presidential and Carter Ranges (Figure 1). The prominent glacially sculpted Carter Notch is also seen in this direction. An ice-marginal lake was impounded in the north-sloping Peabody Valley during one or more episodes of glacial advance and retreat (see Stop 1, Day 3). Striations in the Peabody Valley were originally interpreted as indicating northeastward flow of alpine glacial ice from the Presidential Range (Vose, 1868). However, stoss-and-lee erosional features show that the actual flow direction was toward the southwest, as continental ice was deflected up the valley by the local topography (Hitchcock, 1878).
To the west, we see the upper reach of the Moose River valley, which forms part of a glacial trough called "Randolph Valley". This valley straddles the divide between the Connecticut River basin to the west and the Androscoggin basin to the east. It played an important role in the late Wisconsinan glacial history of the northern White Mountains. Striations indicate that Randolph Valley channeled continental ice from the Connecticut Valley lobe eastward around the north end of the Presidential Range (Goldthwait, 1940). Ice flowing through this valley also may have contributed to deposition of the Androscoggin Moraine (Stop 1; Thompson and Fowler, in press). During deglaciation, meltwater from ice remaining in the Connecticut River basin poured through Randolph Valley into the Gorham area (Stop 2; Gerath, 1978). Meltwater channels and ice-contact deposits were formed as the glacial tongue thinned and retreated westward from the drainage divide at Bowman, in the highest part of the valley (Goldthwait and Mickelson, 1982). These ice-contact deposits are at least partly deltaic, suggesting that a glacial lake briefly existed between the ice margin and the Bowman divide (Thompson and Fowler, in press).

To the north of Randolph Valley is the Crescent Range. From Pine Mountain we can see a deep meltwater channel (Icy Gulch) cutting across this range northeast of Mt. Crescent. (This feature may be observed more easily with a low sun angle and absence of leaf cover.) The Icy Gulch channel was cut early in the deglacial history of the area, during the "nunatak phase" (Goldthwait and Mickelson, 1982).

The view northeast from Pine Mountain shows the Androscoggin River valley. The valley turns abruptly eastward at Gorham and becomes very narrow where it passes between the Carter and Mahoosuc Ranges. This constricted portion of the Androscoggin Valley was an outlet for glacial ice impinging on the northwest flanks of the adjacent mountain ranges. An ice stream flowing down the valley formed the moraine system seen at Stop 1.

STOP 4 (B. K. Fowler, P. T. Davis, R. P. Goldthwait) - Durand Lake Deposits, Randolph

At this stop we will examine some controversial sediments in the vicinity of Durand Lake, on U.S. Route 2 west of Gorham. Gravelly deposits at the mouths of several narrow valleys draining the northern Presidential Range contain stones derived from higher parts of the valleys to the south. Bradley (1981) interpreted these deposits as till emplaced by alpine glaciers issuing from cirques in the Presidential Range during late-glacial time. Fowler (1984) subsequently presented evidence that the Durand Lake deposits resulted from colluvial processes, rather than glacial transport from the King Ravine cirque. The latter view is shared by Davis and Waitt (1986), who also have mapped this area. Waitt and Davis (1988) summarized the arguments and data pertaining to the sequence of cirque glaciers and continental ice in the Presidential Range, as well as on Mt. Katahdin in Maine and the Green Mountains in Vermont. Reprints of Fowler's (1984) and Waitt and Davis' (1988) articles will be provided to field trip participants.
STOP 5 (R. F. Gerath) – Glacial Features of the Gorham-Milan Area

The specific location(s) of this stop will be determined just prior to the trip, based on availability of exposures. We will examine features produced by the late Wisconsinan ice sheet in the upper Androscoggin River valley, between Gorham and Milan (Figure 5). The glacial geology of this area was mapped at 1:24,000 scale by Gerath (1978) and summarized by Gerath, Fowler, and Haselton (1985). We will note the contrast between the rugged terrain around Gorham and the open aspect of the Androscoggin Valley north of Berlin. There is about 1,000 m of relief where the Androscoggin River cuts through the mountains east of Gorham.

There are many glacially streamlined features with low relief (~100 m) in the unincorporated township of Success, New Hampshire. The Mahoosuc Range rises abruptly 400 m above the Success lowland (Figure 5). Ice-formed topography in the Success area, and striations and grooves in the Mahoosucs, indicate that the Laurentide Ice Sheet flowed southeastward over this portion of the White Mountains. The ice sheet was probably on the order of 1,700 m thick at the late-Wisconsinan glacial maximum roughly 18,000 yr ago. The radiocarbon chronology of southern Quebec and Maine, and other evidence (Gerath, 1978), indicate that this portion of the Androscoggin Valley was deglaciated no later than about 12,600 yr B.P.

The vertical ablation of over 1,000 m of Laurentide ice left a sparse record that includes glacial spillways in high cols such as Mahoosuc Notch and the Bear Spring Brook meltwater channel (Stop 2; Figure 5). Sequences of stratified drift finally formed in areas where reduced topographic gradients allowed the deposits to be preserved.

Important aspects of the local deglaciation are shown in the valley profiles compiled by Gerath (1978). The river gradients above Berlin are controlled by bedrock at the heads of gorges which drop 70 m over a 4 km reach formerly known as Berlin Falls. This important ice and water-scoured valley sector accounts for 23% of the elevation range (300 m) of all the stratified drift in the area. The splendor of Berlin Falls is now subdued by hydroelectric works.

Gerath (1978) described the deglaciation of the upper Androscoggin River basin. Thick glacial ice persisted in the nearby Connecticut River basin until the deglaciation of the Androscoggin Valley was well advanced. Two ice fronts retreated northwestward in the Dead and Androscoggin Valleys above Berlin after ice in the lower valley was beheaded at the falls. Much of the stratified drift below Berlin was deposited in contact with dead ice. Gerath has interpreted several lines of evidence to propose that all of the stratified drift in the 300 m of elevation between the northwest foot of the Mahoosucs and the floor of the Androscoggin Valley formed in about two centuries.

Steep slopes and valley gradients in the mountains, and continually lowering and retreating ice surfaces, resulted in stratified drift being reworked and deposited at progressively lower elevations. Extensive depositional gradients were only preserved at the final stage of deglaciation, when base levels were controlled by isolated masses of dead ice or bedrock on valley floors.
Figure 5. Map of glacial features in the Milan-Berlin-Gorham area. Map explanation is on facing page. From Gerath and others (1985).
MAP LEGEND
(Elevations in meters a.s.l.)
STRATIFIED DRIFT AND ALLUVIUM

Limits of Stratified Drift and Alluvial Deposits in the Androscoggin River and Copperville Through Valleys. Annotations refer to deposit descriptions given in Table 1. Area beyond valley bottoms consists of till, colluvium and bedrock. Limits of stratified deposits in the major valleys of the Success area are unmapped.

MELTWATER EROSIONAL OR DEPOSITIONAL FEATURES

Esker. Direction of depositional flow known or inferred.

Drift - Incised Meltwater Channel. Abandoned Meltwater channels at least partly incised in drift.

Sequentially Occupied Meltwater Spillways. Variously defined abandoned meltwater channels draining directly into the Androscoggin Valley from the Success Area. Approximate elevation of threshold.

Sequential Ice - Marginal Channels. Ice-controlled channels partly incised in drift by meltwater issuing from the Moose River Valley.

Head of Ice - Contact Stratified Drift. Marks head of depositional sequences in Copperville Through Valley and upper Androscoggin Valley.

GLACIAL ICE EROSIONAL FEATURES

Drumlinoid Hill. Glacially moulded and streamlined bedrock-controlled landform with variable elliptical shape.


Longitudinal Till Ridge. Very elongate till ridge. These ridges occur in groups.

BEDROCK

Valley floor bedrock. Area of extensive outcrop at Berlin Falls (City of Berlin).

Figure 5. Map legend.
Paraglacial depositional environments

The importance of paraglacial conditions was first noted by Church and Ryder (1972), who defined the term as "nonglacial processes that are directly conditioned by glaciation" (p. 3059). The term "paraglacial" is nearly synonymous with Fulton's (1967) Water depositional environment.

Paraglacial environments lingered while there was a large supply of sediment available on recently exposed slopes and regional sources of meltwater were rapidly declining. This was the final transitional phase leading to Holocene fluvial environments. The most distinctive paraglacial features in this area are alluvial fans which are now incised by their parent streams.

Glaciolacustrine sediments

Several areas of glaciolacustrine deposition are mentioned above. A complex drift sequence, including glaciolacustrine deposits, in the Peabody Valley south of Gorham was described by Gosselin (1971) and is currently being investigated by Haselton and Fowler (1988) (Stop 1, Day 3).

The Success moraine is a body of deltaic sand, gravel, and muddy sediments at the foot of the Mahoosuc Range (Figure 5). These deposits were dammed by an ice front as it retreated northward for over 8 km and diverted meltwater into the Androscoggin Valley in a descending series of topographically-controlled lateral spillways (Figures 6 and 7). The Success kame moraine is the highest, and probably the earliest, body of stratified drift deposited during deglaciation of the Milan-Berlin-Shelburne area.
Figure 6. Stereogram showing ice-channel filling deposited by meltwater draining into Cascade-Alpine Brook -- the earliest meltwater spillway from the Success area into the Androscoggin Valley. Note the glacially streamlined features north of Cascade-Alpine Brook. (See Figure 5 for location.) From Gerath (1978).
Figure 7. Stereogram showing lateral drainageways numbered 4 and 5 (see Figure 5) near the mouth of Horne Brook in the Androscoggin Valley. Channel 4 was cut across a north-sloping ridge crest as meltwater from the Success area continued to be diverted by the retreating Androscoggin Valley ice front. From Gerath (1978).
STOP 1 (B. K. Fowler) - Stratigraphy of the lower Peabody River valley, Gorham

Previous investigations

The exposures at this stop (see Location Map), with lacustrine sediments overlain by two diamict units and an intermediate lacustrine unit, have been the source of much study and controversy during the past fifty years. Crosby (1934) cited these exposures, along with a number of others scattered across northern New Hampshire, as evidence of till overlying lacustrine deposits, and suggested they were part of the so-called Bethlehem Moraine. He postulated this moraine to be the result of a late Wisconsinan readvance or stillstand of the Laurentide Ice Sheet. Goldthwait (1958) soundly refuted Crosby's proposition on geomorphological grounds, but later Lougee (1940) and Flint (1953) mentioned the deposits again in support of a proposed late Wisconsinan readvance (perhaps of Valders age).

In 1970, during the 33rd reunion of the Friends of the Pleistocene, the equivocal nature of the stratigraphy at the Peabody River site brought forth a stimulating but inconclusive dialogue concerning the significance of this section. Since that time, and due in part to that discussion, sporadic studies have focused on better stratigraphic description of the section and the search for organic material for dating.

In 1971, Fowler (unpubl. data) examined the rhythmite sequence at the bottom of the westerly exposures and demonstrated that approximately 275 irregular but distinguishable couplets are present in the lower lacustrine unit. Later, Gosselin (1971) published the first detailed description of this section and its multiple exposures.

Goldthwait (1972) performed detailed fabric analyses and stone counts on the two diamict units. His data demonstrated a strong fabric parallel to the valley's axis, and a definite north to northeast provenance for both of the diamict units. This evidence refuted earlier postulations of northward ice flow from a highland ice cap to the south (Flint, 1951) and suggested the deposits were related exclusively to ice moving from the north. Goldthwait (1972) concluded, on the basis of similar fabrics, provenance, and clay mineralogy, that the two diamict units were the result of a single stage of glaciation, with the lower of ablation origin and the upper of colluvial origin (flowtill from ice located to the northeast).

Gerath (1978) studied the geomorphology of these deposits and concluded that they were a morainal "plug" inserted between the steep valley walls, but he considered its origin to be uncertain. He suggested the possibility of pre-Laurentide emplacement, but could not rule out deposition by late Wisconsinan ice to the north in the Androscoggin Valley.

Bradley (1982) also investigated this section and resurrected Crosby's postulation of widespread moraine segments in this region. On the basis of geomorphic observations, he concluded that the Peabody Valley deposits were
formed by ice flowing southward, possibly as a late-glacial ice tongue pushing up the valley and damming a proglacial lake. Gerath and Fowler (1982) refuted Bradley's contention that this and other moraine segments are of regional significance, but did not dispute the importance of the Peabody River section to the regional deglaciation history.

Presently, Fowler and Haselton are conducting detailed stratigraphic description and analysis of the exposures seen at this stop. Their research has become more significant because of work on the nearby Androscoggin Moraine in Shelburne, New Hampshire (Thompson & Fowler, 1986). The possible existence of an ice tongue in the Androscoggin Valley and its relationship to these two localities has renewed the need for an accurate interpretation of this complex section.

Stratigraphic description

The 150-ft thick (46 m) section consists of several distinct units (Figure 1). The basal unit is composed of 70-80 ft (21-24 m) of lacustrine rhythmites and varves. Couplets vary widely in thickness and gradation and suggest a complex, rapidly changing lacustrine environment located close to glacial ice. This unit contains blocks of compact, stony diamict with clasts of northeasterly provenance. Dropstones with conformal bedding surrounding them and scattered thick lenses of fine to coarse sand with graded bedding are also present. It has been reported that this unit lies directly on bedrock at this locality (Goldthwait, pers. comm.), although bedrock is not presently exposed in the riverbed.

The basal unit is overlain by 20-25 ft (6-8 m) of compact, silty, clayey, locally cobbly to bouldery diamict. The contact between these units is sharp; no rip-up clasts or deformation of the contact or the underlying lacustrine unit has been observed. The fabric of the clasts in the diamict is predominately N 40° E (Figure 2a), and stone counts clearly indicate the clasts came from an area lying about 6 miles (10 km) to the north-northeast (Table 1).

The lower diamict unit is overlain by 8-10 ft (3 m) of bouldery/cobbly gravel, which displays little internal bedding or other sedimentary structure except for a locally discontinuous bed of coarse to fine sand near its base. This gravel overlies the diamict along a sharp contact.

Above the lower diamict is a very distinctive 18-20 ft (6 m) sequence of severely deformed, sandy-silty rhythmites and more massive lacustrine beds. Shearing and thickening/thinning of beds has been observed along micro-faulted and deformed surfaces. These structures are best preserved in the more clayey strata. Imbrication of folds and a uniform northerly dip of the micro-faults and sheared structures suggest the deformation was caused by compression from the north. The contact with the underlying gravel is sharp and undeformed, and no rip-up clasts have been observed.

Finally, the deformed lacustrine unit is overlain sharply by 6-10 ft (3 m) of loose, silty, sandy, stony diamict. The fabric of the clasts has a predominate orientation of N 45° E (Figure 2b); clast provenance is identical to that of the lower diamict unit. The top of this unit has an abrupt gradational contact with winnowed, bouldery ground moraine, typical of the surficial deposits in this area.
Figure 1.
Composite Stratigraphic Diagram
Lower Peabody River Valley
Sorham, New Hampshire

Loose Stony to Sandy Diamict
(Upper Till)

Deformed Lacustrine Rhythmite Units
18'-20' thick

Cobbly Gravel
displays little sedimentary structure
8'-10' thick

Compact Cobbly to Silty Diamict
(Lower Till)
20'-25' thick

Lacustrine Rhythmites and Varves containing
Dropstones and Till Black
70'-80' thick
## TABLE 1

Summary of Stone Count Percentages
Lower Peabody River Valley
Gorham, New Hampshire

<table>
<thead>
<tr>
<th>Lithologies@</th>
<th>Lower Diamict Samples</th>
<th>Upper Diamict Samples</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1* 2* 3+ 4+</td>
<td>1* 2* 3+ 4+ 5+ 6+</td>
</tr>
<tr>
<td>Amphibolite</td>
<td>14 12 16 10</td>
<td>12 22 6 6 15 13</td>
</tr>
<tr>
<td>Biotite Granitics</td>
<td>38 40 36 43</td>
<td>34 32 33 22 28 30</td>
</tr>
<tr>
<td>Quartzite</td>
<td>1 4 2 0</td>
<td>6 6 4 0 0 0</td>
</tr>
<tr>
<td>Qtz. Diorite</td>
<td>10 9 15 20</td>
<td>16 17 14 16 18 21</td>
</tr>
<tr>
<td>Qtz. Monzonite</td>
<td>1 0 2 5</td>
<td>3 0 0 0 0 0</td>
</tr>
<tr>
<td>Biotite Granite</td>
<td>29 29 21 18</td>
<td>29 19 21 17 15 18</td>
</tr>
<tr>
<td>Binary Granite</td>
<td>4 1 6 4</td>
<td>0 1 14 35 24 18</td>
</tr>
<tr>
<td>Biotite Schist</td>
<td>0 0 0 0</td>
<td>0 2 1 0 0 0</td>
</tr>
<tr>
<td>Undifferentiated</td>
<td>3 5 2 0</td>
<td>0 1 7 4 0 0</td>
</tr>
<tr>
<td>Totals</td>
<td>100 100 100 100</td>
<td>100 100 100 100 100 100</td>
</tr>
</tbody>
</table>

Notes: @ Assemblage derived from area north-northeast of Peabody River section, dominated by widespread outcrops of biotite granites and granitic gneisses of the Oliverian Plutonic Series, which have been extensively intruded by biotite granite and quartz diorite with some secondary quartz monzonite of the White Mountain Plutonic Series. Rock types to the south consist primarily of sillimanite-grade, aluminous metasedimentary rocks (primarily mica schist and para-gneiss with minor amounts of quartzite).

* Count percentages summarized from Goldthwait, 1972.

+ Unpublished count percentages of Fowler.
Figure 2A.
Composite Rose Diagram
Clast Orientation, Lower Diamict Unit
Lower Peabody River Valley
Gorham, New Hampshire

Note: Diagram based upon 150 clast measurements from 3 counts: 64, Goldthwait (1972); 50, Fowler (unpubl.); and 36, Haselton (unpubl.)
Figure 2B.
Composite Rose Diagram
Clast Orientation, Upper Diamict Unit
Lower Peabody River Valley
Gorham, New Hampshire

Note: Diagram based upon 92 clast measurements
from 2 counts: 42, Goldthwait (1972); 50,
Fowler (unpubl.)
Status of Interpretation

The interpretation below is preliminary, because field and interpretive work have not been completed. At this stage, results suggest that the Peabody River section represents various types of deposition in an ice-contact proglacial lake.

Ashley has characterized the sedimentary environment in such a lake as complex because of the types of deposition possible from interactions between glacially-derived detritus, solid and melting ice, and water. Deposits range from unsorted, bouldery till to well-sorted glaciolacustrine silts and clays, occurring in an almost infinite variety of lateral and vertical combinations. Sedimentary events are usually rapid and separated by long periods of quiescence. Diurnal and seasonal melting create large discharges with high concentrations of sediment during relatively short periods of time, so that the bulk of the sedimentation at a site may take place during only a fraction of the time that ice is receding from the area (Ashley, 1985).

The particular combination (or combinations) of ice, water, detritus, and time that occur at a site will determine its stratigraphy. For example, rapid melting of buried ice masses creates locally-steep slopes of unsorted saturated detritus. These conditions lead to debris-slope failures, transport by turbid flow, and redeposition of detritus as diamict of varying textures and stratigraphic descriptions (Shaw, 1985; see also Hartshorn, 1958; Boulton, 1968; and Lawson, 1979). If this process alternates with episodes of slower melting and transport by traction and suspension, deposits are created with interbedded diamict and stratified units (Persson, 1983; Lawson, 1982).

Presently, interpretations of the Peabody River section suggest that the alternating diamict and lacustrine units formed in the complex sedimentary environment described above. It is likely this environment included steep gradients, due to local topography and wasting ice. Plentiful supplies of glacially derived detritus were deposited by runoff events, ranging from high-volume, turbid flows to lower-volume, fine sediment flows.

Davis and others (1980) have shown that the local climate alternated between periods of warmth with heavy runoff and cold periods with hard freezes of long duration. In this climate, alternating sedimentary units could have been deposited rapidly during the warm periods of meltwater discharge, and could then have been consolidated and perhaps frozen during the following period of cold quiescence. These units have sharp contacts and lack of grading and incorporation between them. This suggests their emplacement during independent episodes, over the possibly frozen surface of the previously deposited unit. The repetitive character of the section (lacustrine-diamict-lacustrine-diamict) appears more likely due to dynamic fluctuation in this sedimentary environment, than to widespread late Wisconsinan events.

Based upon the stratigraphy and the foregoing environmental interpretations, it appears that these deposits are not necessarily the result of a widespread regional readvance of late Wisconsinan ice of the sort contemplated by Crosby and Bradley. They do suggest local readvance of ice in the Androscoggin Valley, pushing southward to this site and damming at least one proglacial lake.
The lowland extending from the town of Conway in the Saco River valley to Ossipee in the Ossipee River valley contains numerous erosional and depositional features associated with Pleistocene glaciation (Newton, 1974). The area lies at the edge of the Lakes Region, bounded on the north by the White Mountains. This topographic setting greatly influenced the flow of ice through the area and resulted in a number of deeply scoured valleys. Perhaps the best example of this is the portion of the Saco River valley extending northward from Conway to the village of Intervale. The valley here appears to have been overdeepened by coalescing ice streams flowing out of Crawford and Pinkham Notches. These ice streams trimmed the valley walls to form Cathedral and White Horse Ledges near the village of North Conway.

Meltwater channels and meltwater deposits suggest that deglaciation in this region was controlled by the underlying topography. The ridges acted as barriers separating areas of active ice to the north from stagnant ice to the south. The through valleys were occupied by tongues of ice which received tremendous amounts of meltwater and sediment from the melting ice to the north. Most of this sediment was transported through the valleys and out onto the outwash plains to the south.

Although the steep east faces of Cathedral and White Horse Ledges appear to be parallel to the regional ice flow in the Saco River valley in the North Conway area, Ellis (1972) considered the cliffs to be lee sides of classic roche moutonnées. The ledges are composed of Conway Granite, which was intruded as coalescing stocks at the eastern edge of the White Mountain batholith. They are part of the White Mountain plutonic-volcanic series, dating about 185 ma. Although the granite crystallized at depth, volcanism is evidenced by flows, tuffs, and breccias of the Moat Volcanics on Moat Mountain, a few kilometers southwest of the ledges.

The Conway Granite is a coarse, pink biotite granite that generally weathers to a dull gray. The granite is not foliated, but in a few places schlieren with oriented quartz, biotite, and feldspar grains may be found. Miarolitic cavities lined with smoky quartz and feldspar crystals are locally abundant.

The Conway Granite is generally massive with joints commonly tens of meters apart. Ellis (1972) summarized a detailed study that he made of the two types of joints. Primary jointing formed during deep burial soon after the magma crystallized. Based on the presence or absence of quartz, aplite, or basalt dikes, Ellis (1972) distinguished two ages of near-vertical primary joints with variable orientations. Sheeting, as is well displayed by exfoliation slabs on White Horse Ledge, formed due to unloading of the rock by erosion. Sheets of rock on the tops and sides of the ledges are common and less than a meter thick; however, sheets are less common and much thicker with depth at the bases of the ledges. The sheeting is younger than the primary jointing because exfoliation joints cut across the primary joints, dikes, and schlieren. Also, the sheeting is preglacial in origin because exfoliation joints have been partially removed by glacial erosion.
On the stoss (northwest) sides of the ledges, sheeting is generally intact, as glacial ice only abraded the bedrock surfaces. Grooves, crescentic gouges, polish, and perched boulders are common. However, on lee (southeast) sides of the ledges, sheeting is truncated by the cliff faces (east side of Cathedral Ledge and south side of White Horse Ledge), as evidenced by some sheets overhanging space created by removal of bedrock by glacial plucking. Much of this erosion may have occurred when continental ice was thick enough to have flowed from the northwest. Later, when continental ice thinned, the topography controlled a generally southward ice flow, which polished bedrock surfaces, such as the slabs of White Horse Ledge. Very little evidence exists for development of sheeting due to glacial unloading, except perhaps in a few places on the slabs of White Horse Ledge. Thus, Ellis (1972) suggested that a longer period of time than the postglacial is required to develop the extensive sheeting observed on Cathedral and White Horse Ledges and in other quarries in New England.

Note: Because of the generally unweathered character of the Conway Granite, along with the variety of cracks that follow the vertical primary jointing and the moderate-angled exfoliation slabs that follow the sheeting, Cathedral and White Horse Ledges are one of the most popular rock-climbing areas in North America. In recent years, hang-gliding from the tops of the ledges also has become common.

STOP 3 (R. M. Newton) - Government Pit, Albany

At Conway the Saco River valley makes a right-angle turn to the east. However, much of the glacial ice continued streaming southward, eroding a series of "through valleys" across the divide between the Saco River and Ossipee River watersheds (Figure 3). These through valleys range up to 13 km in length and 100 m in depth. The surrounding hills also show strong evidence of glacial erosion as many have a well-developed stoss-and-lee form comparable to roche moutonees.

Not all the bedrock in the region was glacially scoured down to fresh, solid rock. There are some exposures of deeply weathered granite on the lee sides of hills, which appear to have been protected from glacial erosion. The weathered granite is mined as it makes excellent road metal. The resulting exposures frequently reveal sets of vertical and horizontal clastic dikes which have been injected into the granite, presumably during glaciation. Similar features can also be found in some of the exposures of pre-late Wisconsinan till.

The Government Pit shows excellent exposures of disintegrated Conway Granite (rottenstone). The granite is weathered to a depth of at least 6 m at this locality. Weathering is concentrated along joint planes, and in places has resulted in rottenstone surrounding cores of relatively fresh bedrock. Till overlies the rottenstone in part of the pit area (Newton, 1977).

The Government Pit is also one of the best known mineral localities in the state. Collectors have honeycombed the weathered rock in search of miarolitic cavities containing crystals of smoky quartz, feldspar, topaz, and rare beryllium minerals.
Figure 3. Map of the Ossipee Lake 15-minute quadrangle, showing the locations of through valleys. From Newton (1974).
STOP 4 (R. M. Newton) - Bald Hill meltwater channel, Albany

Meltwater channels occur on steeper, higher elevation slopes than the meltwater deposits. Some of the channels may be associated with the progressive melting of stagnant ice, while others appear to have formed between stagnant and active ice. A set of eight parallel lateral channels on the side of Mt. Chocorua range in elevation from 200 m to 400 m. These channels appear to have been formed due to the progressive thinning of stagnant ice occupying the adjacent valley. Another set is found in the area of Bald Hill, just west of the village of Conway. Here two channels were cut through adjacent gaps in an east-west ridge and can be traced 3 km downslope to kame terraces in the Silver Lake through valley (Figure 4). For meltwater to have flowed through this system the ice at the north end of the channel must have stood at an elevation of at least 300 m, while at the downstream end it appears to have been at an elevation of only 180 m. This suggests a rather steep gradient for the ice front (22 m/km).

STOP 5 (R. M. Newton) - Kame terrace-esker complex, Rte. 113, Madison

There are extensive deposits of stratified drift throughout the lowland areas. The through valleys are filled with ice-contact deposits which feed into extensive areas of outwash farther south. In some cases the ice-contact deposits may be directly correlated to meltwater erosion features at higher elevations. The ice-contact deposits include kames, kame terraces, and eskers. The eskers occur primarily along the bottoms of the through valleys. In at least one case the meltwater stream which formed the esker can be shown to be hydrostatically controlled so that the flow direction was opposite to the topographic gradient.

STOP 6 (R. M. Newton) - Outwash plain, near Silver Lake, Madison

The largest outwash plain lies in the area between Ossipee and Silver Lakes and covers an area of approximately 15 km² (Figure 5). The outwash grades from coarse gravel on the north to medium sand to the south. Much of the southern part of the fan is deltaic. There are a series of kame deltas south of the outwash fan which indicate an early high-level stage of the glacial lake into which the outwash built.

STOP 7 (R. M. Newton) - "Lower Till", Rte. 25, Tamworth

Most of the area is covered by a loose, sandy till with numerous cobbles and boulders. Some of the boulders lying on the surface are quite large; the largest is the "Madison Boulder" measuring 25 m x 11 m x 7 m. This till correlates with the "Upper Till" found throughout most of New England (Koteff and Pessl, 1985). A second, older till outcrops sporadically throughout the area, with the most numerous exposures within the Ossipee Mountains. The till seen at this stop correlates with the "Lower Till" found in southern New England and here averages 57 percent sand, 21 percent silt, and 22 percent clay. Exposures in the Ossipee Mountains reveal both a brown, weathered "oxidized" zone and an olive-gray unweathered zone. There are both vertical and horizontal sets of clastic dikes cutting through many of the exposures.
Figure 4. Map showing the locations of the col channels west of Conway. The Sugar Loaf channel formed first, followed by the Chase Hill channel. From Newton (1974).
Figure 5. Map of the maximum extent of the Silver Lake outwash fan. S and G denote outcrops of sand and gravel, respectively. Arrows indicate the flow directions of source streams onto the outwash fan. From Newton (1974).
Most of the larger lakes in central New Hampshire lie in bedrock basins 15 to 60 m deep, scoured out by repeated glacial erosion of weaker and deeply rotted rock types (granitic ?). They are surrounded by discontinuous hills and low mountains (schists, quartzites, and volcanics) that reach elevations up to 450 m higher than the lakes.

As the last Wisconsinan ice sheet disappeared from this area 13,000 to 14,000 years ago, it left abundant evidence of meltwater flow in the form of channels, kame terraces, and high deltas. The ice did not disappear simultaneously from all lake basins; nor did it retreat with one uniform steep edge across basin to basin. Rather it thinned first in one basin and then in another as shown by meltwater flow. These basin sequences are illustrated by details in the Ossipee-Wolfeboro-Winnipesaukee area. The order of opening of the basins from thick ice lenses to open lakes was generally south to north and east to west.

Where broad and deep north-south valleys, like the Merrimack and Connecticut River valleys, penetrate the hills and mountains, the residual but thinning ice sources in northern New Hampshire and Vermont kept broad streams of ice flowing into these valleys long after surrounding basins were deglaciated (12,000 yr B.P.?). This is shown by the steep ice-contact features (Merrimack River valley, sequential outwashes) and ice-contact deltas (Connecticut River valley, glacial Lake Hitchcock). These areas opened up from south to north as demonstrated by the work of others.

The final stage of ice retreat from central New Hampshire was melting of ice from valley bottoms (Goldthwait and Mickelson, 1982). The deepest parts of valleys held the thickest ice, hence were the last to become deglaciated. Ice-marginal lakes developed locally where ice blocked the surface drainage. Drainage also occurred around and over separated ice masses, leaving terraces with ice-contact faces. When the Lake Wentworth basin first became ice-free, water could not drain to the southwest into Lake Winnipesaukee, so a lake formed which overflowed to the southeast (Goldthwait, 1968). A large sand delta with a lobate front, foreset beds, and a flat top near the overflow threshold, 21 m above the present lake, is evidence for a former glacial lake (Figure 6). Therefore, the northern part of the Lake Winnipesaukee basin was under ice when the Lake Wentworth basin opened up (Goldthwait and Mickelson, 1982).
Figure 6. Channels crossing shoulder of hill near Wolfeboro, and chain of ice-contact deposits formed by meltwater flowing eastward into glacial Lake Wentworth. K, kame; d, delta; s, sand plain; m, till and bedrock hills. From Goldthwait and Mickelson (1982).
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INTRODUCTION

This trip is intended to show the origin and early history of glacial Lake Hitchcock (figs. 1 and 2), describe some of the major events of deglaciation that occurred during retreat of the Laurentide ice sheet in the southern part of the Connecticut River valley, and indicate what the lake and post-lake features suggest to us about the nature of postglacial uplift. We have benefited greatly from a vast amount of work that has been done in this region in the last several decades, which is now being compiled for the new State surficial geologic maps of Connecticut and Massachusetts. These compilations are based primarily on detailed geologic mapping at 1/2-minute scale, and most of these maps have been published by the U.S. Geological Survey or the Connecticut Geological and Natural History Survey; there are also many field-trip guides, theses, and other reports available. A separate investigation of the nature of postglacial uplift has been conducted recently, which also has taken advantage of the detailed mapping.

Our present understanding of Lake Hitchcock has benefited not only from the detailed mapping, but especially from ideas and local stratigraphic details that emerged from the state-map compilations. Many observations and conclusions that seemed contradictory in the past, even from one large-scale map to another, appear to have been successfully resolved at a regional scale. Also, the more concentrated work on postglacial uplift data has provided a model that helps explain quite a few of the earlier inconsistencies. Even though the modern work has allowed a more integrated concept for the deglaciation and postglacial uplift of the Connecticut Valley, there remain far too many other unanswered questions. Some of these questions are the subject of this trip. It should be stressed that much of how we presently view things was anticipated by several workers in the past.

We are indebted to Phil Schafer and Byron Stone who spent much effort in helping us put this guidebook together. Some of their ideas have been incorporated here and they have assisted in reviewing the manuscript.
Figure 1. Glacial Lake Hitchcock and delta deposits in Massachusetts and Connecticut. Ice-marginal deltas record high lake levels as far north as Chicopee, Mass. Northward from there to as far north as Burke, Vermont, ice-marginal deltas record the stable lake level. All stable-level deltas south of Chicopee were built by meteoric water from major tributary valleys such as the Westfield, Chicopee, Farmington and Scantic River valleys.
Figure 1. Generalized outline of glacial Lake Hitchcock and selected other glacial lake areas in western New England. (N) glacial Lake Nashua; (S) glacial Lake Sudbury. (••••••) location of altitude obtained from unmodified, ice-marginal or meltwater-derived delta used in regression analysis described in the text. Uplift isobase interval 25 m. Figure from Koteff and Larsen (in press).
EARLY STUDIES

Glacial Lake Hitchcock, which is now thought to have extended well over 200 miles (320 km) from central Connecticut to Burke, Vt. (fig. 2), was given its name by R. J. Lougee (1939) because of Edward Hitchcock's (1818) mention of evidence for lake deposits between the town of Gill and Mt. Holyoke, Mass. Hitchcock's description was somewhat brief and it seems clear that no glacial source was assumed. The name for the lake became firmly established in the 1950's and 1960's during detailed quadrangle studies in Connecticut and Massachusetts where it was accepted by a number of workers.

B. K. Emerson (1898a,b) thought that the glacial sediments in the Connecticut Valley required the presence of ponded water, but he seems to have viewed the lake as more of a "tremendously swollen stream." He gave the names Springfield Lake, Hadley Lake, and Montague Lake for separate areas, mostly in Massachusetts. Later, Emerson (1917) also recognized the effects of postglacial uplift in the region. He stated that "The lakes are bordered by a bench, which is well marked where it cuts into sand beds or drumlins and broadens in great delta flats at the mouth of tributary valleys," and that "As there was almost no southward current in these lakes the beach (bench) must have been nearly horizontal, and the basin in the northern part of the State must subsequently have been elevated nearly 200 feet more than on the south line."

In a paper on the clays and clay industries of Connecticut, G. F. Loughlin (1905) was the first to have recognized several of the most important aspects of Lake Hitchcock that are still valid today. Remarkably, in three rather short paragraphs, he identified the "kames and high gravel plain" at Rocky Hill, Conn., as the dam across the Connecticut River valley, the outlet for the lake near Newington Station (now called the New Britain channel), and a water level at the outlet at or a little above 80 feet. Loughlin also recognized that the southward sloping clay deposits north of the outlet were the result of "depression of the continent to the northward at that time," which is an obvious reference to postglacial uplift.

R. F. Flint (1933) referred to the lacustrine deposits in Connecticut as belonging to the "Hartford lake," continuing the perception that the Connecticut Valley contained several separate glacial lakes. He also referred to the outlet as "the channel at New Britain," although most of the feature, including the apparent threshold, is in Newington. It is not clear why Flint chose the name New Britain in favor of Newington, but his description has been accepted and used for over fifty years.

Lougee (1939), in naming Lake Hitchcock, considered it to be one integrated body of water. However, he thought it extended farther south than now placed. He did not recognize the outlet at New Britain, as it was never mentioned in his publications. R. H. Jahns and M. E. Willard (1942) specifically addressed the previous notion of separate water bodies in their detailed analysis of the Massachusetts portion of the lake, demonstrating that the lake features defined a single lake, the level of which was controlled by the New Britain channel. This work was done during the initial stages of the detailed mapping program in Massachusetts, and many of their ideas and descriptions of Lake Hitchcock features have been altered only slightly by later studies. Some of Jahns' concepts of sequences and systematic ice retreat were developed here at this time, although he did not include deltas in this original scheme.

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No attempt has been made in this all too brief discussion of the early work to cover all the important contributors (for example Antevs and his varve chronology in 1922, and Flint's first 1930 study). Although these are only a few highlights of how the lake history was first established, it is clear that Lake Hitchcock has been the object of much interest and study dating back to the last century. Also, there have been many workers in recent decades who have contributed a great amount of detail to the geology of the lake, many of whom will be referred to in the discussions at the field stops.

**INITIATION OF GLACIAL LAKE HITCHCOCK**

The inception of Lake Hitchcock was dependent on the presence of an earlier and higher glacial lake, Lake Middletown (Stone et al., 1982) (fig. 3), in the Connecticut River valley at Middletown and the tributary Mattabesset River valley during retreat of the Connecticut Valley lobe of the Laurentide ice sheet. An extensive deltaic complex controlled by the level of Lake Middletown completely filled the Connecticut River valley in the vicinity of Rocky Hill. This sediment mass later formed the dam for Lake Hitchcock. Formation of such drift dams in south-draining valleys has been found to be a necessary condition for the creation of many glacial lakes in southern New England. Lake Middletown itself was impounded by a long mass of older meltwater sediments that effectively filled the lower Connecticut River valley southeast of Middletown. Because these deposits extended at least 12 mi (20 km) down the valley, entrenchment of them and consequent lowering of Lake Middletown was relatively slow.

Construction of the deltaic complex began with deposition of successive, contiguous ice-marginal deltas in Lake Middletown in the Cromwell area, which blocked that relatively narrow part of the Connecticut River valley (see STOP 1, fig.5). As the ice margin retreated from the Cromwell deltas, meltwater was impounded behind them at a very slightly higher level than Lake Middletown, and ice-marginal deltas formed in this higher lake near Rocky Hill and on the east side of the Connecticut River in Glastonbury. The waters of this relatively small lake spilled over the Cromwell deposits. A well-developed channel (see STOP 1, figs. 5 and 6), called the Dividend Brook Spillway (Hartshorn and Koteff, 1968) was carved into the Cromwell delta surface. Together, the Cromwell-Rocky Hill-Glastonbury deltas are referred to as the drift dam at Rocky Hill.

Ice retreat during the formation of the deltaic complex uncovered, west of the Connecticut River, a bedrock upland that now forms the east-west divide between two tributaries of the Connecticut River, the Mattabesset River to the south and the Park River to the north (fig. 3). The small lake controlled by the Dividend Brook spillway expanded northward behind the deltaic complex, east of Cedar Mountain. At the same time, Lake Middletown expanded northward from the Mattabesset basin across a low part of the divide in the New Britain-Newington area, west of Cedar Mountain. As ice retreated from the north end of Cedar Mountain, the lake level behind the delta complex dropped and the lake coalesced with Lake Middletown. The Dividend Brook spillway was abandoned and erosion of its channel ceased. The final floor altitude of this spillway was 129 ft, controlled by the level of Lake Middletown into which it drained. The deltaic complex therefore survived to constitute the dam for Lake Hitchcock.
Figure 3. Extent of glacial Lake Middletown.
Although Lake Middletown continued to lower slowly by entrenchment of its drift dam, it remained high enough to cover the low part of the divide (no higher than 110 ft (34 m)) in the New Britain-Newington area and the lake was able to expand northward into the Connecticut River basin during ice retreat. Altitudes of deltas on both the east and west side of the basin indicate that Lake Middletown persisted, but with slowly lowering levels, until the ice margin retreated as far north as Windsor.

Further lowering of Lake Middletown allowed emergence of the low divide area at New Britain-Newington, and separated the shrinking Middletown lake to the south and the first phase of glacial Lake Hitchcock to the north. As the ice margin in the Connecticut Valley retreated northward, Lake Hitchcock expanded in area although its level gradually lowered because of erosion of till, waterlaid sediments, and weathered bedrock in the spillway. Meltwater-fed deltas were constructed successively northward in the lake during stagnation-zone retreat, and their lowering altitudes northward reflect the erosion of the drift at the New Britain channel area. This period of lowering, referred to here as the Connecticut phase of Lake Hitchcock, lasted until the floor of the New Britain channel stabilized on resistant bedrock, preventing further lowering of the lake level, and initiating the stable phase of Lake Hitchcock. By this time, the ice margin may have been as far north as Chicopee, Mass., but its exact position is still unclear.

During its stable phase, Lake Hitchcock continued to expand as the ice margin retreated north from Chicopee through all of Massachusetts and much of New Hampshire and Vermont. Meltwater-fed deltas were built successively in the lake probably to about Burke, Vt. The stagnation-zone retreat of the margin was generally systematic, interrupted in places by local readvances such as one at Chicopee (Larsen, 1982). Most of these readvances have been identified only in recent years, and no doubt others will be found as new exposures become available. However, none of them is believed to represent more than local and short-lived events and thus are not correlated regionally.

Lake Hitchcock was once thought to have drained catastrophically when the ice margin had reached just north of Hanover, N.H. (Lougee, 1939, 1957). Recent work by Koteff and Larsen (1985, in prep.) on postglacial uplift studies, however, has established the longer lake to Burke; also, the presence of post-lake stream terraces along the Connecticut River at the Cromwell-Rocky Hill-Glastonbury drift dam only about 30 feet (10 m) below the projected level of Lake Hitchcock indicates a somewhat less dramatic end to the lake.

Although we still lack definitive evidence to place absolute ages on the inception and the demise of glacial Lake Hitchcock, its life-span can be placed in a regional time-frame. From a regional array of radiocarbon dates and correlated ice-marginal deposits, Stone and Borns (1986) have placed the ice margin in the southern part of the lake basin at 16,000–17,000 years B.P. Varve correlation studies (Antevs, 1922) indicate a minimum of 4000 years for the life of the lake. Koteff and Larsen (1985), emphasizing radiocarbon dates reported by Davis and Ford (1982) from the White Mountains area of New Hampshire, suggest that Lake Hitchcock was still in existence, with its level controlled by the New Britain spillway, at 14,000 years B.P. Larsen (1984) provided evidence for drainage of Lake Hitchcock before 12,500 years ago, and perhaps even before 13,000 years ago while the ice sheet remained in the Winooski River valley of northern Vermont. A radiocarbon date of 12,200± 350 years B.P. was reported by Colton (1960)
from wood in a small peat bog excavated during runway construction at Bradley Airport, Windsor Locks, CT (see fig. 11). Previously, little significance as to the time of lake drainage was placed on the 12,200 year date. It now appears from our regional compilation that this dated locality lies on the lake-bottom surface in front of the Bradley Field delta, and that the date thus records a time after the lake had drained. A radiocarbon date of 10,710± 330 years B.P. (Preston, Person, and Deevey, 1955) from transported wood fragments in gravel at the mouth of the New Britain spillway channel (fig. 7) was thought by Flint (1956) to represent a time when Lake Hitchcock still spilled through this channel. Persistence of the lake until this 10,710 year date seems unlikely in view of recent evidence; re-evaluation of the stratigraphic horizon from which this dated material was taken is still needed.

POSTGLACIAL UPLIFT

The deposits of Glacial Lake Hitchcock present an unusual opportunity for uplift studies. The lake was lengthy (more than 200 miles [320 km]), lasted for at least 4000 years with a stable outlet for probably half that time, and was located in an area that was deglaciated early. Also, the physical correlation and relative position of most of the deposits are well known because of the detailed mapping of much of the lake area, and we have been able to identify a large number of ice-marginal or meltwater-fed deltas that were deposited successively in Lake Hitchcock during systematic ice retreat. Altitudes obtained from topset/foreset contacts in these deltas now record the postglacial tilt of a once-level water plane.

Although Loughlin (1905) and Emerson (1917) early on suggested that the area had undergone postglacial uplift, it was Lougee (1939, 1957), who first did any detailed studies. He carefully surveyed altitudes of topset/foreset contacts of Lake Hitchcock deltas and from these reported uplift gradients to the north-northwest of 3.3 ft/mi (0.63 m/km) for Connecticut (and presumably Massachusetts as well), and of 4.6 ft/mi (0.87 m/km) for New Hampshire. Jahns and Willard (1942) also used altitudes of topset/foreset contacts of deltas in Massachusetts and determined the uplift gradient there to be approximately 4.2 ft/mi. Recent studies by Koteff and Larsen (1985, in prep.), used similar techniques and have arrived at slightly different conclusions. From the recent studies, the uplift gradient indicated for the entire area covered by Lake Hitchcock from central Connecticut to northern New Hampshire and Vermont is 4.74 ft/mi to the N 20-21 W (fig. 4).

This uplift gradient was established by examination of more than 60 delta localities in Massachusetts, New Hampshire, and Vermont. Delta localities in Connecticut were not included initially because of the complex history of a gradually lowering lake there; by using deltas north of there associated only with the stable phase, a constantly changing variable was excluded from the study. However, some of the Connecticut delta features are addressed on this fieldtrip. Of the 60 delta localities, 28 were selected as representing unmodified deltas resulting from successive meltwater deposition at the ice margin in Lake Hitchcock or from meltwater streams that entered the lake from tributary valleys. The others were considered to be modified by collapse or erosion by later meteoric water, or were constructed in later and lower lake levels after either uplift began or the drift dam at Cromwell-Rocky Hill-Glastonbury, Conn., failed.
Figure 4. Ordinary least squares regression profile based on altitudes of topset/foreset contacts of 28 unmodified, ice-marginal or meltwater-derived deltas (+) in glacial Lake Hitchcock. (•) other altitudinal data. Dashed profiles diagrammatic only. Lake-bottom profile estimated from previous publications and topographic maps; lake bottom may be higher at delta localities 7 and 8 (STOP 5 discussion). Figure from Koteff and Larsen (in press).
Topset/foreset contacts (T/F) of deltas are a very consistent and accurate estimate of former glacial lake levels, probably to within 3 ft (1 m). This principle has been known for many years (Gilbert, 1890, fig. 15). In our study, deep erosional fluvial channels were avoided; in many of the deltas, the topset beds are 3 ft (1 m) thick or less over foresets. Thus, the water-level error due to erosional scour at the T/F is minimal. Most of T/F altitudes were surveyed with a transit, alidade, or electronic distance meter. In most cases, a permanent bench mark was used for control; in a few other cases, road intersections with elevations located to the nearest foot were used so that the T/F altitude is accurate to within that amount. A few altitudes reported by Jahns and Willard (1942) were used and the accuracy of them is less certain because they did not describe their field methods. However, these altitudes were field checked and found to be reasonable.

Most of the T/F altitudes (fluvial/foreset contacts in some cases) are shown on figure 4. The profile was originally derived however, from altitudes of only the 28 unmodified meltwater-fed deltas mentioned earlier because they represent the stable level of Lake Hitchcock during deglaciation (our attitude about a few of these at the southern end of the profile has been modified in putting together this trip, to our benefit obviously, and are discussed at the field stops). There is a vertical difference in uplift between the lake spillway at New Britain and the northernmost delta in Vermont of 720 ft (219 m), over a distance of about 152 mi (245 km). The gradient of the profile is thus 4.74 ft/mi (0.9 m/km).

The profile is a best-fit projection based on an ordinary least squares regression of the 28 T/F altitudes. The regression indicates a N20°1/2-21W direction for the projection with error range for the E-W variable of 5% and 0.4% for the N-S variable. Two sigma variation for each altitude is less than 6 ft (2 m). Only two of the delta altitudes are more than 6 ft (2 m) off the fit (one of these, at Chicopee, Mass., may actually represent the last part of the higher Connecticut phase of Lake Hitchcock), and 22 of the altitudes are within 3 ft (1 m). Projection of the profile southward to the lake spillway indicates that the threshold of stable Lake Hitchcock was about 82 ft (25 m) altitude. Drilling supervised by J. W. Bingham of the USGS Water Resources Division, Hartford, indicates that the bedrock floor at the threshold is about 58 ft (17.7 m) altitude. The water column there is indicated to have been about 24 ft (7 m) in a channel about 700 ft (215 m) wide, and the discharge rate for the lake is calculated to have been about 215,000 ft³/s (6100 m³/s). Only two modern floods in the basin covered by Lake Hitchcock, recorded in 1936 and 1938, have exceeded this discharge rate, so it seems reasonable that the New Britain spillway could have handled a body of water the size of Lake Hitchcock.

Some of the altitudes reported by Jahns and Willard do not fit well on their generally northward projection of uplift, but do so on the N20°1/2-21W projection. Also, some of the deltas examined by them have now been determined to be later features and not constructed during ice-marginal retreat. Thus, the gradient of 4.2 ft/mi (0.8 m/km), which is an average of all of their data points, is clearly too low. The 3.3 ft/mi (0.63 m/km) uplift gradient reported for Connecticut (and presumably Massachusetts) by Lougee (1939) is no doubt the result of placing the threshold for Lake Hitchcock much farther south than the New Britain spillway. He believed that the uplift projection was about N15W, from which he derived an uplift gradient for New Hampshire of 4.6 ft/mi (0.87 m/km), reasonably close to that of Koteff and Larsen (1985, in press). Lougee explained the different gradients as the result of a hinge line. A N15W projection from the New Britain channel area, however, produces a similar uplift gradient of about 4.6 ft/mi (0.87 m/km) for Connecticut, Massachusetts, and New Hampshire. There is no need to employ hinge lines to describe uplift in this region.
DISCUSSION

The nature of the uplift profile (fig. 4) for the Connecticut Valley indicates that the style of postglacial rebound in this region is significantly different than that derived from water bodies in other areas, particularly those that were deglaciated later. The straightness of the uplift profile and the extraordinary closeness of fit of the regression show absolutely no differential warping of the lithosphere. Also, rather than being a time line, the profile is a time-transgressive depiction of ice-marginal or near ice-marginal delta construction in Lake Hitchcock during a systematically northward but increasing rate of ice retreat. As inferred from the correlation of Stone and Borns (1986), the retreating ice margin was in the vicinity of Chicopee, Mass., the southernmost delta locality used for the profile, between 15,000 and 16,000 years ago. Koteff and Larsen (1985, in press) place the ice margin at the northernmost delta about 14,000 years ago. Thus, the profile represents between 1500 and 2000 years of ice retreat. During this time and possibly longer, the stable phase of Lake Hitchcock was maintained at a constant level by the bedrock-floored spillway at New Britain. All of this suggests that postglacial uplift was delayed until the ice was at least in northern New England about 14,000 B.P. If postglacial uplift was delayed during this period of ice retreat that lasted 1500-2000 years, it is further suggested that uplift was delayed from the beginning of ice retreat from Long Island more than 19,500 years ago (Sirkin, 1982) as well. It seems unlikely that uplift could have been occurring in southern Connecticut and Long Island without affecting any part of glacial Lake Hitchcock during deglaciation there. The entire region appears to have been affected by uplift only after 14,000 B.P., when the ice margin is assumed to be in northern New England. Depending on dating accuracy, a delayed response to uplift of about 5000 years is proposed, from the beginning of ice retreat from Long Island until the ice margin was in northern New Hampshire and Vermont.

Another style of postglacial uplift has been suggested by J. A. Clark (in prep.) that depicts active uplift at the ice margin from the beginning of deglaciation. Among other things, this model assumes that ice retreat was fairly steady. However, as indicated by Stone and Borns (1986) and Schafer (1979), it is probable that the rate of ice retreat was twice as fast over New Hampshire and Vermont as it was over Connecticut and Massachusetts. Indeed, the rate of retreat may have increased gradually even from the ice position at New Britain. Also, several readvance localities are known in the Connecticut Valley, particularly the southern part, suggesting that ice retreat really was not very steady, although it certainly was systematic. Clark's model projects a series of time lines from each delta point to the spillway which fall below the straight profile shown in figure 4. Although we can not show Clark's profile adequately here, it is discussed at various places during the trip, particularly at the Chicopee delta. Clark's model produces a convex up profile, which is about 20 ft (6 m) off the straight-line projection near the center (Groen, Clark, and Koteff, 1986). However, the straightness of the projection (fig. 4) derived from precisely determined delta altitudes seems to preclude a convex up or any curved depiction.

There no doubt are other models of postglacial uplift that differ from the suggestion here that there was a significant delay to the uplift response at the beginning of deglaciation. But it should be stressed that this area is the only one so far that has been studied carefully in a region deglaciated early, before 14,000 B.P. All data for other postglacial uplift studies have been derived from areas that were deglaciated after 14,000 B.P. It is hoped that there is evidence here to provoke a healthy discussion.
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ARCHAEOLOGICAL PERSPECTIVES ON HOLOCENE FLUVIAL RATES AND PROCESSES IN THE CONNECTICUT RIVER VALLEY

Dena F. Dincauze

Archaeological research can contribute crucial data to geomorphological and sedimentological investigations. Archaeological sites can be dated by a variety of geochronological methods, and once the chronology is established, archaeological cross-dating using short-lived artifact styles can extend it to sites and areas where none of the basic geochronological methods can be applied. Furthermore, archaeological stylistic dating, properly applied, can provide ages close to the precision of radiocarbon.

In areas such as the Connecticut River Valley, where river dynamics during the Holocene have removed and emplaced large amounts of sediment, collaboration between archaeologists and geologists can be highly productive. Archaeological materials occurring on geomorphological surfaces or in sediment bodies can date those surfaces and bodies with a precision exceeding that which geologists typically expect. Sites on surfaces provide terminus ante quem ages for those surfaces; artifacts included in sediment bodies and sites or artifacts stratified beneath them provide terminus post quem ages. When archaeological survey and inventory activities have been intense in an area, such bracketing dates can be both reliable and precise. The identification of Paleoindian sites, dating to the 11th century BP, on alluvial and aeolian surfaces within the bounds of Lake Hitchcock provided one of the first solid indications that the drainage date for Glacial Lake Hitchcock needed to be revised backward (Curran and Dincauze 1977).

In the Hadley stretch of river, and farther north in Deerfield, near Turner's Falls and in Northfield, archaeological site sequences on post-lake terraces provide indications of varying rates and discontinuous episodes of fluvial erosion and deposition during the Holocene. At this provisional stage of research, some of these observations appear to require modifications in the fluvial history of the area. Artifact-style chronologies may ultimately help to correlate discontinuous terraces along stretches of the river; they certainly discriminate late Holocene terraces near the modern flood plain. Archaeology may also help to bracket the periods during which major aeolian bodies were deposited, by indicating ages for terraces with and without major dunes, and by their stratigraphical relationships to the extensive sand sheets and dune fields on the lake floor. In the Connecticut stretch of the river, alluvial bodies have been found enclosing archaeological sites; to date, only equivocal indications of such relationships have been found in Massachusetts.

During trip B1, archaeological sites that illustrate some of these issues will be visited, as the route and timing of the itinerary permit. Discussion and the posing of alternative interpretations are anticipated with pleasure.

References:

Curran, Mary Lou, and Dena F. Dincauze
FIELD TRIP STOPS
(see Figure 1 for general locations)

STOP 1. MUSTARD BOWL PITS; town of Rocky Hill Conn., Hartford South quadrangle.
Turn east from Main St., (Rte. 99) 0.3 mi (0.5 km) south of Cromwell-Rocky Hill town line, onto unimproved road and travel approximately 1 mi (1.6 km) to end of road at southeast corner of southernmost pit scarp (fig. 5).

The pit access road crosses part of the surface of the Cromwell-Rocky Hill-Glastonbury delta complex, which is a series of ice-marginal deltas that completely filled the Connecticut River valley between Rocky Hill and Glastonbury to an altitude of 150-180 ft (46-49 m). This mass of deposits provided the dam for glacial Lake Hitchcock after Lake Middletown had lowered (see text discussion). The dam is now entrenched by the Connecticut River. Inset against the higher surface is a terrace remnant at 50 ft (15 m) altitude, which was cut probably at the time the dam was breached and Lake Hitchcock drained.

The Cromwell-Rocky Hill-Glastonbury delta complex consists of deposits controlled by two water planes (figs. 5 and 3). The older southern deltas were built into open water of Lake Middletown and completely blocked the valley at highest altitudes of 160-170 ft (49-52 m). When the ice margin retreated slightly, but still impinging against Cedar Mountain to the northwest, meltwater was ponded behind the heads of the Lake Middletown deltas and spilled across them through a well-developed channel that straddles the Cromwell-Rocky Hill town line just east of Rte. 3 (figs. 5 and 6). This channel, called the Dividend Brook spillway (Hartshorn and Koteff, 1968), was the base-level control for several sequential ice-marginal deltaic deposits that make up the northern part of the Cromwell-Rocky Hill-Glastonbury complex. The spillway was carved into the delta surface from about 150 ft (46 m) down to its present floor altitude of 129 ft (39 m). Deepening of the channel was controlled by the presence of Lake Middletown at its mouth, which had lowered to just below 130 ft (39 m) by the time drainage through the spillway ceased.

The Mustard Bowl pits are cut into the first delta controlled by the Dividend Brook spillway. The topset/foreset contact exposed in this delta is estimated to be 146-149 ft (44-45 m) in altitude. North of the Mustard Bowl kettle and east of Dividend Pond (fig. 5), several pit faces expose about 100 ft (30 m) of ice-marginal and deltaic sediments. At the ice contact northeast part of the deposit, coarse-grained severely collapsed ice-marginal deposits are excavated in the lower pit in the floor of the main pit. The north-facing scarp exposes proximal, interbedded gravel and sand foreset beds on the east and, pebbly sand foreset and bottomset beds to the west. In the lower foreset beds, fine to medium sand beds include ripple-drift cross-laminated units and associated draped lamination, interbedded with planar beds. In the middle to upper foreset beds, pebbly sand, pebbly gravel, medium to coarse sand and silty sand beds dipping 10-15 degrees to the southwest show planar beds and megaripples in transverse bed forms. Fluvial gravel topset beds are exposed best in the farthest west scarps above the 150-ft (46-m) contour. The topset bed sequence is 10-12 ft (3-4 m) thick. The pit centered on the Mustard Bowl kettle shows gentle collapse of delta topset and foreset beds toward the center of the kettle. The surface of the isolated ice block that produced the kettle was at least partly below lake level. By the end of deposition, the ice block was mostly or completely buried by delta sediments derived from meltwater streams issuing from the main ice mass to the northeast.
Figure 5. Cromwell-Rocky Hill-Glastonbury delta complex which formed the drift dam for glacial Lake Hitchcock. Base from Hartford South and Glastonbury 7½' topographic quadrangles, 1984 editions.
Figure 6. 1970 aerial photograph of Dividend Brook spillway. Compare with Figure 5.
STOP 2. NEW BRITAIN CHANNEL, SPILLWAY FOR GLACIAL LAKE HITCHCOCK;
towns of Newington and New Britain, Conn., New Britain and Hartford South
quadrangles. Parts of the channel can be seen from several localities; we will stop
at the north end, on the west side. This point can be reached by traveling west on
Rte. 174 (New Britain Ave.), turn right (north) on Charles St., just west of the
Newington-New Britain townline; travel about 0.3 mi (0.5 km), turn right (east) on
Judd Ave., travel 0.1 mi (0.16 km), bear right on 8th St., which curves northward,
travel about 0.5 mi (0.8 km) and park on the right at end of 8th St., and junction with
Conant St.

The narrowest part of the New Britain channel is overlooked from our viewpoint; it
is at the head of the clearly erosional part of the spillway and is about 700 ft (213 m)
wide. Drilling conducted by the Water Resources Division, USGS, Hartford, in the
channel floor at the narrows showed sandstone and mudstone at 58 ft (17.7 m) altitude.
The narrows acted as the threshold for the stable phase of Lake Hitchcock at a projected
water level (see text) of 82 ft (25 m) altitude. Discharge capacity at this level at the
narrows is calculated to have been 215,000 ft³/s (6100 m³/s). The flat-floored, partly
steep-walled channel (figs. 7 and 8) extends about 2.2 mi (3.5 km) south from the narrows
to its mouth near the Newington-Berlin town line; it is about 900-1400 ft (274-427 m)
wide. Crumbly, easily eroded sandstone and siltstone are exposed at several places on
the channel walls. Till is locally thick and at some places overlies stratified sand or
rhythmites, perhaps of pre-late-Wisconsinan age. An ice-marginal delta built into Lake
Middletown forms the west wall of the channel north of the narrows, and was terraced
during downcutting to the stable channel level.

The spillway crosses the east-west divide between the basins of two tributaries of
the Connecticut River: Mattabesset River to the south and Park River to the north. The
divide area near the spillway includes NNE-trending ridges of Jurassic sedimentary rocks
and basalt. The conspicuous topographic discordance between the channel and the
adjacent upland topography shows that the channel is an erosional feature formed after
the retreat of the last ice sheet (see fig. 8). The topography of the upland adjacent to
the channel narrows indicates that the pre-erosion altitude of the divide at this site
probably was no higher than 110 ft (34 m), and perhaps even lower. This is well below the
possible altitude of Lake Middletown (+/- 150 ft; 46 m), which flooded the divide area as
a strait initially about 0.6 mi (1 km) wide. As the level of Lake Middletown lowered, the
strait became narrower and shallower. One may guess that when the depth of water in
the strait lowered sufficiently, a riffle and a slight southward gradient developed on the
water surface. This change marks the separation from Lake Middletown of the water
body north of the divide and the inception of Lake Hitchcock at its earliest and highest
level.

About a mile (1.6 km) east of the New Britain channel narrows, another low point
on the Mattabesset-Park divide occurs at about 85 ft (26 m) altitude at the head of
Rockhole Brook. In contrast to the strikingly erosional form of the New Britain channel,
the sag at the head of Rockhole Brook shows no indication on topographic maps, air
photos, or on the ground, of water erosion; erosional forms downstream along the Brook
are appropriate to the Brook itself. The basin of Rockhole Brook is largely occupied by
severely collapsed ice-marginal deltas graded to Lake Middletown, with surface altitudes
of 150-170 ft (46-52 m). Evidently, this delta complex together with extensive bodies of
dead ice sufficiently blocked the Rockhole Brook sag, preventing the spillway of Lake
Hitchcock from forming there.
Figure 7. New Britain channel (floor shown by shaded area), spillway for glacial Lake Hitchcock. New Britain and Hartford South 7½' topographic quadrangles, 1946 ed. ▲ indicates approximate site of 10,710±350 BP date (Preston et al, 1955).
Figure 8. Stereo-paired aerial photographs of the New Britain channel spillway area; photographs taken in 1970. Compare with Figure 7.
STOP 3. KELSEY-FERGUSON CLAY PIT; town of South Windsor, Manchester quadrangle. From junction of Rt. 291 (Bissell Bridge) with Rt. 5, proceed 2.7 mi (1.7 km) north on Rt. 5. Turn right at light onto Strong Rd., then immediate left onto Brickwood Lane. Turn right into Kelsey-Ferguson Brick Co.

Stream terrace deposits, 10-20 ft (3-6 m) thick, overlie lake-bottom deposits of Lake Hitchcock, as much as 175 ft (53 m) thick, in the vicinity of STOP 3 (see fig. 9). Sand dunes up to 30 ft (9 m) high occur on the terrace surface. The stream terrace deposits, composed predominantly of sand, were formed during and after the draining of Lake Hitchcock as the drift dam at Rocky Hill was breached and entrenched. Surface altitudes of the terrace are 75-85 ft in the area of STOP 3. The terrace deposits are inset only slightly, if at all, into the lake-bottom surface which is at 70-80 ft altitude in this area. Lake waters dropped about 40 ft (12 m) before terrace sediments were deposited.

Figure 9. Part of the Geologic map of the Manchester quadrangle (Colton, 1965) in the vicinity of STOP 3. Q1 - lake-bottom deposits, Qst - stream terrace deposits, Qsd - sand dune deposits, Qal - floodplain alluvium, Qt - till, TRp - Portland Arkose. Scale 1:24,000.
This exposure is in the only active clay pit in the Lake Hitchcock basin and provides a rare glimpse into the bottom sediments. The clay pit is located two miles from the lake margin and removed from major point sources of glacially derived sediment. The lake at time of deposition was 55-70 ft (17-21 m) deep. Sediments are rhythmically bedded dark clays and light-colored multiple graded silts (fig. 10). However, there is enough inconsistency in the bedding (i.e. thin clay bands in silt and thin silt layers within clay) to present problems in interpreting these rhythmites as varves. It is difficult to determine which silt laminae represent the commencement of a yearly deposit and which are the result of a chance introduction of silt into the lake during clay deposition. It is likely that sediments were transported to the site by overflows and/or interflows.

Figure 10. Rhythmically bedded silt and clay, Kelsey-Ferguson clay pit, South Windsor, CT.
STOP 4. FARMINGTON RIVER DELTA COMPLEX; town of Windsor, Windsor Locks quadrangle.  a) GREAT POND DELTA PIT; Turn north off Prospect Hill Road, 0.15 mi (0.24 km) east of junction with Blue Hills Ave. (Conn. Rte. 187). Travel 0.5 mi (0.8 km) to tobacco barns; continue on dirt road past barns to pit.  b) WINDSOR DELTA PIT; Turn north off Prospect Hill Road 1.0 mi (1.6 km) northeast of junction with Day Hill Road, onto Long Road; Turn left onto new development road about 1500 ft from Long Road junction.

The Farmington River delta complex fans out to the north and south in the lower reaches of the Farmington River (fig. 11). The separate deltas of the complex were built into successively lower lake levels and resulted from progradation over a relatively long period of time, from ice-marginal deposition when lake levels were high (Connecticut phase), through Farmington River meteoric deposition during a long stable phase, to a short post-stable phase.

The Great Pond delta (fig. 11) is the earliest and highest level delta in the Farmington River complex. It prograded southwestward from a NW-trending ice-margin position which marks the west side of the Connecticut Valley ice lobe (figs. 1 and 2). At the time of Great Pond delta deposition, Lake Middletow still covered the New Britain Channel area but at a considerably lowered level; probably just prior to the emergence of the New Britain-Newington divide and the initiation of Lake Hitchcock as a separate lake.

The Windsor delta (fig. 11) was built from an ice margin position about a mile northeast of the head of the Great Pond delta. Progradation of the northern and eastern parts of the delta from the ice margin was in part synchronous with deposition of the western and southern part by distal meltwater entering the lake through the Tariffville Gap (fig. 11). A topset/foreset contact at 178.6 ft measured in the ice-marginal part of the delta (near Stop 4b) projected to the New Britain channel records a water-level of 117 ft at the spillway threshold. This level, with allowance for a modest depth of water over the spillway, indicates that erosion of the initial 110-115 ft land surface had begun.

The Bradley Field delta (fig. 11) was built northeastward into the lake by water entering the lake through the Tariffville Gap after the ice-margin had retreated from the area north of the present Farmington River. Although deposition in this area probably began while lake levels were still high, the extensive Bradley Field delta was constructed chiefly during the long stable phase of Lake Hitchcock. Altitudes of topset/foreset contacts seen at several construction sites lie on the projected stable-phase lake line. A topset/lacustrine sand contact exposed by backhoe excavation in the northeast part of the delta was surveyed by Stone, Koteff, and Stone at 154 ft; this altitude falls on the stable-phase lake line.

The last depositional event recorded in the Farmington delta complex is an erosional terrace, inset slightly into the Bradley Field delta. The terrace is on grade southeastward to delta plains north and south of Farmington River, here called the Kennedy Road delta deposits (fig. 11). These surfaces protrude southeastward from the rest of the delta complex; the fluvial plain of the southern part of this delta is about 10 ft below the Lake Hitchcock stable phase water plane. The base of fluvial sediments in the surface north of the river, estimated at 129 ft altitude, falls 7-8 ft below the stable lake level. The relative lowering of lake level recorded by these deposits probably
Figure 11. Farmington River delta complex. ▲ indicates locality of 12,200±350 B.P. date (Colton, 1960).
indicates that either entrenchment of the dam or postglacial tilting had begun at the time of their construction. When the Rocky Hill dam failed and Lake Hitchcock drained the Farmington River cut deeply through the delta complex.

Both pits (4a and b) expose an upward-coarsening sequence of deltaic beds. At the base, several meters of fine-to-medium sand contain laterally extensive beds that are subhorizontal, but in places show dips of as much as 10°. Vertical sequences of ripple-drift cross-laminations and draped laminations of white fine sand and red silt record waxing and waning density underflows that flowed on the shallow lake bottom in front of the prograding edge of the delta alluvial plain. In a few places in the pits, sandy foreset beds dip more than 15°-20°. Coarse, pebbly sand, in festooned trough cross-beds, disconformably overlie the lacustrine sand beds. The trough cross-beds are overlain by planar-tabular cross-beds of coarse, pebbly sand, and interbedded thin beds of pebble gravel. The trough cross-beds and related gravel beds are interpreted as a coarsening-upward glaciofluvial sequence, related to the prograding braided alluvial plain of the delta.

STOP 5. SCANTIC RIVER DELTA COMPLEX, POWDER HILL ROAD PIT; town of Enfield Broad Brook quadrangle. From I-91, interchange 47, travel eastward on Hazard Ave. (Rte. 190) 2.4 mi (4 km) through village of Hazardville; turn right (south) on Powder Hill Road; cross Scantic River (rapids over bedrock on left); pit described by Ashley, et al., (1982) on left immediately across River. Continue on Powder Hill Road 0.5 mi (0.83 km) to large pit on left.

Like the Farmington River delta complex, the Scantic River complex records lowering levels of the Connecticut phase and the stable phase of Lake Hitchcock. The early ice-marginal delta south of the Scantic River with surface altitude of 200 ft (fig. 12) was built in front of a NE-trending ice margin into a high level of Lake Hitchcock. A delta surface that reaches 190 ft altitude lies mostly north of the river. Subaqueous beds of this delta, however, extend south of the river along Powder Hill Road and are exposed in the upper section of the Stop 5 pit. This non-ice-marginal delta is at the distal end of a fluvial meltwater terrace which has its ice-marginal head in the Scantic River valley in Massachusetts; it was built into Lake Hitchcock, still in Connecticut phase, but slightly lower than the 200-ft ice-marginal delta. Younger and lower 150-160-ft surfaces lie on the western margin of the delta complex (fig. 12). These surfaces are at the stable-phase Lake Hitchcock level; they were constructed by meteoric water from the Scantic River valley and are probably deltaic although at present there are no pits that display the internal deltaic structure. Thin fluvial sand and gravel beds that are graded to these surfaces underlie the 170-ft terrace surface at Stop 5 and the 150-ft terrace surface above the river bank exposure to the north (fig. 12).

The sand and gravel pit exposes superposed and contrasting stratigraphic sections of two deltaic morphosequence units, capped by thin fluvial terrace sediment and eolian sand. The lower unit is well exposed in the lower west pit wall where it comprises a coarsening-upward sequence of beds; from top to bottom:

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Figure 12. Topographic map of the Scantic River delta complex area.
Unit | Thickness | Description
--- | --- | ---
1) | 1-2 m | pebble gravel, with abundant poorly sorted coarse sand matrix, in massive beds. Gravel clasts are red sandstone, basalt, and crystalline rocks; coarse sand contains abundant red sandstone rock fragments. This unit thickens in the southern end of the exposure where thin sand beds and disrupted gravel clast fabric shows probable ice-meltout, collapse deformation.

----- sharp contact -----

2) | 0-0.6 m | flowtill; red, compact, matrix supported diamict sediment; unit is lens-shaped in outcrop, about 6 m long; matrix is silty-sand; clasts are chiefly angular red sandstone.

----- sharp contact -----

3) | 1 m | silt and fine sand beds, interbedded with thin lenses, less than 10 cm thick, of compact sandy red flow-till.

----- covered interval -----

1-2 m | medium-coarse sand and pebbly sand in thinly bedded and laminated foreset beds; which dip south-southeast.

The lower unit extends across the pit floor to the east wall where red compact, flow till is poorly exposed at the base. Similar beds exposed in lower parts of two sections along the river to the north, described by Ashley et al. (1982), indicate that the top of this unit slopes northward and is the collapsed proximal part of the ice-marginal 200-ft delta.

The upper unit is best exposed in the east pit wall, where it is 3-4 m thick. It is chiefly medium to fine sand in horizontal beds containing cosets of climbing-ripple cross-laminations and related draped laminations. The sand is salt and pepper, quartz and dark heavy-minerals with conspicuous biotite. This unit thickens by way of a downward sloping lower contact to the exposure at the south end of the pit. Here, ripple laminations and intrastratal fluid-escape structures are exposed. The lateral continuity of the beds, the stacked vertical sequences of ripple laminations, and the lack of gravel and cross-bedded coarse sand units indicate that these beds are glaciolacustrine in origin, similar to beds in upper part of stream sections to the north (Ashley et al., 1982). These are delta bottomset beds.

Fluvial coarse pebbly sand and thin pebble-gravel beds disconformably overlie the bottomset sandbeds in the upper part of section. Terrace sediments are overlain by gray-buff (oxidized) massive fine sand of eolian origin (Colton, 1965).
The Chicopee delta complex is the largest of those deposited into glacial Lake Hitchcock. Similar to the deltas at the mouths of the Hockanum, Farmington, and Scantic valleys, the Chicopee complex includes ice-marginal deltas, distal meltwater-fed deltas, and possibly meteric deltas. The Chicopee deposits have been divided provisionally (Larsen, F.D., Stone, B.D., personal communication) into four ice-marginal deltaic units (see fig. 1) which trend northeast to southwest and record sedimentation from the east side of the Connecticut Valley ice lobe into lowering lake-levels (Connecticut phase). A measured topset-foreset contact at 225 ft (68.6 m) altitude at Westover Airforce Base, 1.4 mi (2.4 km) northwest of STOP 8, is 8 ft above the stable-lake level (Fig. 4, delta locality 7), and represents one of the last parts of the Connecticut phase of the lake. Distal meltwater from the extensive Chicopee drainage basin then built deltas into the stable level of the lake (see Figure 1) after the margin of the Valley ice lobe had retreated to north of Chicopee. These stable-level deltas are found on the western side of the complex. Meltwater from the Chicopee drainage must have first entrenched earlier ice-marginal deltas before entering the lake, and in some places stable-level deltas probably overlie lacustrine deposits of the earlier ice-marginal deltas. The lake-bottom deposits seen at STOP 6 underlie a stable-level delta surface, but these varved clays accumulated during ice-marginal delta construction, perhaps at the time the delta seen at STOP 8 was built (Larsen, 1982). Evidence for minor readvance of ice in the lake is seen at both STOP 6 and 8. Evidence for decreasing amounts of meltwater-supplied sediment is seen in the lacustrine sections at STOP 7. At STOP 9, fluvial-terrace beds laid down by southward flowing water of the incipient Connecticut River disconformably overlie lacustrine sediments at an altitude of about 205 ft, only 12 ft (3.6 m) below stable-lake level. This section indicates that the surface of lake-bottom sediments was very shallow entirely across the lake in front of the Chicopee delta complex. If there had been active uplift of the area during deglaciation to the north, this shallow lake bottom would have been raised above the level of the lake, and later deltas would have been graded to a new lake level controlled by a spillway across the uplifted lake bottom. To the contrary, the straight-line profile (fig. 4) indicates that ice-marginal deltas to the north were deposited in the stable lake still controlled by the New Britain spillway, and that uplift of the entire region was delayed until the ice margin was in northern New Hampshire and Vermont.

STOP 6. ENTRANCE TO CENTER AUTO PARTS; town of Chicopee, east side of Center Street, 0.25 mi (0.4 km) south of the I-91 bridge over the Connecticut River (fig. 13). When first studied in 1977, the east-west exposure (fig. 13) was 77 ft (24 m) long and up to 16.6 ft (5 m) high. At present, only 10 percent of the original exposure is extant and that is in danger of removal. The section measured in 1977 is as follows:

0.0-5.9 ft (0.0-1.8 m) undisturbed clay-silt varves
5.9-6.1 ft (1.8-1.9 m) brown till
6.1-9.6 ft (1.0-2.9 m) sheared and thrust-faulted varves, minor recumbent folds
9.6-14.7 ft (2.9-4.4 m) grayish-brown till with lenses of cross-bedded pebbly coarse sand
14.7-15.9 ft (4.4-4.8 m) deformed varves: gray silty clay, brown silt, minor brown fine sand
15.9-16.6 ft (4.8-5.1 m) brown till, bottom of till not observed

The section clearly demonstrates readvance of the margin of the Connecticut Valley lobe on the bottom of glacial Lake Hitchcock.
Figure 13. Part of the Springfield North 7½' topographic quadrangle map showing locations of STOPS 6,7,8,9.
STOP 7. CENTER STREET CUTS; town of Chicopee, east side of Center Street just north of the I-91 bridge over the Connecticut River. Two exposures of glaciolacustrine deposits are accessible here (fig. 13).

The southernmost exposure (closest to bridge) is within distal foreset beds of Chicopee delta deposits (Ashley et al., 1982). Water depth at time of sedimentation would have been between 85-100 ft (26-30 m) deep. Sediments are characterized by couplets (i.e. rhythmic bedding) 8-12 cm thick, consisting of medium silt (mean = 8-16μ), multiple grade beds and punctuated by thin graded clay layers. The tops of the clay layers are indicated by arrows on Fig. 14. The incipient ripples below the bottom clay layer and the multiple graded beds suggest that sedimentation occurred by density underflows moving down the shallow slope (<5 deg.) of the delta front.

The northern cut at PARK WRECKING exposes lacustrine beds at 150-190 ft altitude, higher in the section than those in the southern cut. The clay-silt varves exposed are between 1 and 8 in (2.5-20 cm) in thickness.

Figure 14. Rhymically bedded silt and clay, cut just north of I-91 bridge (fig. 13). Arrows indicate tops of clay layers.
STOP 8. ZIELINSKI PIT (formerly BASKIN PIT); town of Chicopee, Springfield North quadrangle, located just northeast of, and adjacent to, Exit 6 of the Massachusetts Turnpike (fig. 13).

There are three features to observe in this pit: (A) brown till up to 45 ft (14 m) thick on the east side of the south wall, (B) low-angle distal foreset beds and proximal bottomsets in an east-facing exposure trending north-south near the middle of the pit, and (C) minor glaciotectonic features associated with the Chicopee readvance at the western end of the pit.

"When first observed in 1977 this pit was less than one-half the size of the present pit. Reddish-brown lodgement till was exposed on the southeast side of the pit. A curved exposure with deltaic beds 9 m high extended southwest, west, and then northwest from the till. Dune bedding in deltaic topsets indicated transport directions between due west and southwest. No evidence of readvance was noted at that time. By June, 1982, the pit had been expanded nearly to its present size, its growth being limited by powerlines. At the western end of the pit were exposed a series of imbricate thrust faults striking N70E and dipping 38° NW. Within the sediments above the thrust faults was a sloping surface marked by pebbles and small lenses of reddish-brown till. I interpret the sloping surface as a gliding plane upon which the margin of eastern sublobe readvanced a short distance. The readvancing ice was relatively clean as it left little debris on the gliding plane when it melted" (Larsen, 1982).

The deltaic beds described above included topsets and foresets (now removed) of a small ice-contact delta that had a surface altitude over 230 ft (70 m). It appears that the delta was built to the southwest between a northeast-southwest-trending ice margin on the northwest and the northeast-southwest-trending till ridge on the southeast. Given its surface altitude and the fact that 1.4 mi (2.4 km) to the northwest we have a measured topset/foreset contact at 225 ft (68.6 m), we can surmise that this delta was built into either a lowering phase of Lake Hitchcock or into a drift-dammed lake. In either case, this delta was not built into low, stable Lake Hitchcock.

At present, minor glaciotectonic features at the west end of the pit are still observable. This site is located within the zone of the Chicopee readvance, a 2.0 to 2.5 mile-wide (3 to 4 km) belt in which exposures of readvance till and other associated glaciotectonic features occur (Larsen, 1982). It is not known whether the ice margin readvanced 2.0 to 2.5 miles (3 to 4 km) or whether it underwent oscillatory retreat through this zone. In either case, the ice margin was that of an active ice lobe that retreated northward in the Connecticut Valley of Connecticut and Massachusetts.

STOP 9. McKINSTRY AVENUE PIT; town of Chicopee, Springfield North quadrangle, pit is located 1.5 miles (2.5 km) N 85 W of Exit 5 of the Massachusetts Turnpike.

At the northwest corner of the pit (fig. 13) are two fresh exposures. On the east 3 to 4 ft (0.9-1.2 m) of pebbly coarse sand overlies 10 feet (3 m) of fine sand with ripple crossbedding dipping to the south. At the exposure on the west 5.5 to 6.5 feet (1.7-2.0 m) of pebbly coarse and medium sand rest disconformably over 8 to 9 feet (2.4-2.7 m) of fine sand with ripple crossbedding dipping to the north. Both planar and trough crossbeds are well displayed in the upper pebbly unit. The average direction of dip from 10 measurements taken in the fluvial crossbeds is S48.5W. The upper pebbly unit is interpreted to be a stream terrace deposit associated with a terrace at 215 ft altitude that extends 1.0 mile (1.6 km) to the east. The lower fine-sand unit is interpreted to be the bottom deposits of Lake Hitchcock. The upper fluvial unit probably represents stream-terrace deposits left by the early (post-Lake Hitchcock) Connecticut River.
STOP 10. MOUNT WARNER DELTA; town of Hadley, Mass., Mt. Holyoke and Mt. Toby quadrangles. From the junction with I-91 in Northampton east on Rte. 9 for 4.5 mi (7.2 km) to junction with Rte. 116 in Hadley north on Rte. 116 for 0.7 mi (1.1 km) to Rocky Hill Rd., west 0.5 mi (0.8 km) and bear right on Huntington Rd.; west on Huntington Rd., 0.9 mi (1.4 km) to Breckinridge Rd., then north 0.1 mi (0.16 m) to pit entrance.

Some of the description of this deposit is taken from W. A. McIvride's masters thesis (1982) at UMASS. The Mt. Warner delta (fig. 15) was deposited directly into Lake Hitchcock from the ice margin and it occupies an area of about 2 mi² (5 km²) and probably averages 250 ft (76 m) thick. Topset beds are composed of yellowish-brown and reddish-brown sand and pebble to cobble gravel; their thickness increases from less than 2 ft (0.6 m) at the distal end of the delta to as much as 6 ft (2 m) near the center.
Bedding in the topsets shows interlayered sand and gravels, cross beds, and scour and fill channel structures. The foreset beds consist of reddish-brown coarse to very coarse sand, with lesser amounts of pebble gravel and fine sand; as much as 50 ft (15 m) of foresets have been exposed. They dip west to southwest from 20 to 25 degrees and even steeper. In the past, the distal foreset slope has been exposed. The northeast facing slope of the delta marks a former ice-marginal position and collapsed beds have been observed there in places. Possible flowtill has been reported by McIlvride to be interlayered with the topsets, but at present time, we are not sure if this exposure is still available. A topset/foreset contact at 278 ft (84.7 m) altitude obtained by Koteff and Larsen from the Mt. Warner delta falls 1 ft (0.3 m) above the profile; in the Florence Street delta in Northampton along the same uplift isobase 5.4 mi (8.6 km) to the WSW, Larsen has obtained a topset/foreset contact altitude of 277 ft (84.4 m), which is exactly on the profile.

STOP 11. LONG PLAIN DELTA; town of Sunderland, Mass., Mt. Toby quadrangle. From Rte. 116 turn right (east) on Bull Hill Road. Travel 1.4 mi (2.3 km) up frontal slope and across surface of the delta. Turn right (south) on Rte. 63, travel about 1.0 mi (1.6 km), turn right (west) onto dirt road which goes to 4 tobacco barns and pit on left.

The Long Plain delta (fig. 16) is a non-ice-marginal delta, fed by meltwater streams that flowed from the ice margin at the head of Long Plain Brook Valley just east of Mt. Toby. It is an ideal morphologic example of a coarse-grained delta, characterized by steeply dipping gravel and sand foreset beds. The surveyed altitude of the T/F contact, 295 ft (89.9 m), is on the projected line of the stable phase of Lake Hitchcock. Jahns (1951) mapped erosional shoreline benches at the 295-ft contour interval on the north and south ends of the delta front. The surface slope of the delta fluvial plain is .0043 (22.5 ft per mile). Morphologic elements of braided-stream channels are preserved on the delta surface in minor topographic relief of less than 10 ft.
Figure 16. Topography of Long Plain delta, STOP 11.
STOP 12. BENNETTS BROOK DELTA; town of Northfield, Mass., Northfield quadrangle. From the junction with I-91 in Bernardston, Mass., east on Rte. 10 approximately 3 mi (4.8 km) to Rte. 142, Gill Rd.; north for a little more than 1 mi (1.6 km) on Gill Rd., to the general area of large gravel pit operations (fig. 17). Excavation at this pit is sporadic, but most accessible part is east of the road.

Figure 17. Topography of Bennetts Brook delta, STOP 12.
The Bennetts Brook Delta is located on the west side of the Connecticut River approximately 2 km west of the village of Northfield, Massachusetts. In the early 1970's a south-facing pit wall revealed overlapping lens-shaped sand bodies (Fig. A1). The sand bodies were approximately 75 m long and 10 m high and plunged southward under the floor of the pit. Numerous sequences of ripple-drift cross-lamination had migrated up the west sides of the lenses and down the east sides with little change in the thickness of the ripple-drift units. Paleocurrent directions measured across one of the sand bodies show that turbidity currents flowed eastward approximately parallel to contour lines and in some cases actually flowed upslope (Fig. A2). There is no evidence of structural deformation, thus attitude of the sediments must be considered as primary. The sequences were deposited across previously existing topography, perhaps southward-sloping delta lobes.

Figure A1. Three south-dipping sand lobes in the Bennetts Brook Delta
The dominant sedimentary structure within the sequences is ripple-drift: type A (erosional stoss), type B (depositional stoss), and draped lamination. The type produced is related to the relative importance of rate of ripple migration and the rate of bed aggradation (Fig. A3). The ripples climb at some angle \( \theta \) whose tangent is the mean aggradation rate \( V_y \) divided by the downstream migration \( V_x \).

Figure A3.
A measured section 190 cm long taken near the crest of the sand lobe shows type B ripple-drift cross-lamination (Fig. A4). The draped lamination contains a few incipient ripples suggesting periods when bed-load transport was renewed. A slightly deformed clay lamina occurs near the top of the draped lamination. Type A ripple-drift cross-lamination appears next and grades upward into type B, which in turn shows an increasing angle of climb and grades upward into draped lamination. The second unit of draped lamination grades upward into type B ripple-drift cross-lamination, which in turn gives way to type A. This unit of type A grades upward into type B, which, with a gradually increasing angle of climb, grades upward into draped lamination. A thin clay lamina occurs near the top of the parallel lamination.

Flume studies were carried out at the Hydraulic Laboratory at M.I.T. in order to: (1) reproduce some of the characteristic vertical successions of structures found in natural climbing ripple sequences (Fig. A4) and put constraints on parameters (mainly current velocity, rate of aggradation, and time) important in determining the nature of the structures. Figure A5 depicts a comparison of a flume generated (A) and natural (B) ripple drift sequence.
FIGURE A5.

(A) Flume run 8. The run was designed to last 200 min with an asymmetrical velocity curve (maximum velocity = 25 cm sec⁻¹) and a symmetrical aggradation-rate curve. Total accumulation was 18 cm. Flow was from left to right. Starting with a train of ripples that had reached equilibrium with an earlier, stronger flow, deposition began (at arrow) with draped lamination (DL) followed by Type A (erosional-stoss) cross-lamination (A), Type B (depositional stoss) cross-lamination (B), and a final blanket of draped lamination.

(B) A climbing ripple sequence exposed in a glaciolacustrine delta (Bennett’s Brook Delta, glacial Lake Hitchcock, Massachusetts, USA) exhibits a sequence of sedimentary structures similar to that produced in run 8. Flow was from left to right. The sequence begins by deposition over draped lamination (at arrow). Type A climbing-ripple cross-lamination grades into Type B and finally into draped lamination at the top.
GLACIAL AND DEGLACIAL LANDFORMS OF THE AMHERST AREA, NORTH-CENTRAL MASSACHUSETTS

by

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and
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INTRODUCTION

The deglaciation of western Massachusetts was characterized by stagnation zone retreat that produced a variety of ice-contact landforms and glacial lacustrine deposits. Since the development of the morphosequence concept as defined by Jahns (1941, 1953) and his contemporaries at the U.S. Geological Survey, the surficial geology of Massachusetts and much of New England has been interpreted and mapped as a series of retreating glacial fluvial systems each graded from a temporary position of the ice to local base levels controlled by topographic highs or proglacial lakes (Koteff, 1974). For many years it has been envisioned that much of the debris in the glacial fluvial deposits was derived from surface meltwater eroding supraglacial debris derived from shear zones at the outer margin between active and stagnant ice (Koteff, 1974). This "dirt machine" concept has recently been challenged by Gustavson and Boothroyd (1987) who argue that much of this debris was derived from subglacial mass movement and fluvial erosion via a network of subglacial and englacial tunnels. Based upon analogies with the modern Malaspina Glacier, Alaska (Hartshorn, 1952), they argue that portions of the New England sector of the Laurentide Ice Sheet had a similar hydrologic system. The role of subglacial meltwater in the erosion and deposition of subglacial bedforms has also become a common theme in other sectors of the Ice Sheet (Shaw, 1983; Shaw and Kvill, 1984; Sharpe, 1985; Shaw and Sharpe, 1987).

The purpose of this field trip is to highlight some of the more common landforms associated with the deglaciation of the Connecticut River Valley. These landforms, drumlins, ice-contact deltas, meltwater deltas, eskers, kame terraces, post-glacial river terraces, lake floor plains, dunes, and kettles form a surficial mosaic typical of many north-south trending structural valleys in central New England. Intimately related to deglaciation in this area is the northward expansion of Glacial Lake Hitchcock which provided a base level for ice marginal and englacial drainage.

Much of what we know about the deglacial history in the northcentral portion of the Connecticut River Valley is based upon
years of surficial mapping at a scale of 1:24,000 (7 1/2 minute quadrangles) by the U.S. Geological Survey and others (Segerstrom, 1955a, 1955b, 1959; Balk, 1959; Jahns, 1951, 1966; Caggiano, 1977; Campbell and Hartshorn, 1980; among others). Emerson (1898a, 1898b) and Antevs (1922) were among the first to study the glacial lake clays in the valley. Subsequent work by Lougee (1940), Currier (1941), and Jahns and Willard (1942) outlined the sequential retreat of late Pleistocene ice in the region. It was at this time that Jahns (1953) developed Currier's ideas of stagnation-zone retreat into a detailed chronology of deglaciation based upon the mapping of retreating outwash sequences.

**DATING OF DEGLACIATION**

Deglaciation of the northcentral Connecticut River Valley is dated indirectly by inference and extrapolation from radiocarbon dates at sites outside the area. Based upon a regional assessment of radiocarbon-dated materials along the southern margin, Stone and Borns (1986) suggest that the Laurentide Ice Sheet reached its maximum extent along the Ronkonkoma-Beacon Hill-Martha's Vineyard-Nantucket end moraine at sometime between 21 and 18 ka BP (see page 171, this volume, for regional map). Dates bracketing recessional positions, such as those marked by the Charlestown, Buzzards Bay, and Sandwich moraines, suggest that deglaciation proceeded from west to east toward the Great South Channel (Schafer and Hartshorn, 1965).

Stone and Borns (1986) place the ice margin in central Connecticut at the south end of Glacial Lake Hitchcock at about 16,000 to 17,000 BP. The oldest dates from the Champlain Sea suggest that the Laurentide Ice Sheet had retreated north of the St. Lawrence River by about 12 to 12.5 ka BP, the error dependent upon whether dates on wood from Lake Iroquois or marine shells from the lacustrine-marine transition are considered (See Stone and Borns, 1986, for discussion). At a minimum, this leaves about 3500 years for the ice to retreat northward through New England. By extrapolation, Koteff et al. (page 170, this volume) place the ice margin at the Chicopee Delta of Glacial Lake Hitchcock near Springfield, MA, at 15 to 16 ka BP, followed by retreat to the most northern kame delta near Burke, VT, by 14 ka. The literal interpretation of these bracketing dates suggests that deglaciation through north-central Massachusetts probably occurred over a period of less than 1000 years between 14 and 15 ka BP; by inference, many of the landforms visited during this field trip were probably deposited and developed over a period of less than 10-100 years.

**STYLE OF DEGLACIATION**

Jahns (1953), among others, recognized that northward deglaciation throughout much of New England proceeded by continuous downwasting and thinning of the ice surface. In areas of moderate relief, ice first thinned and retreated over the highlands leaving a tongue of active to slowly stagnating ice projecting further down.
valley from the main ice front. From the active ice zone toward the proglacial environment, ice-contact and glacial fluvial deposits formed a graded continuum dispersing sediments away from the ice. As the ice margin continued to retreat, a new head of outwash was occupied and graded to adjusting base levels downstream. Jahns (1953) and later Koteff (1974) developed these observations into what is now widely known as the morphosequence concept. Koteff's primary contribution to the concept was to outline types of morphosequences that differ depending upon ice contact relationships and the proximity of proglacial lakes as a local base level control (Fig. 1).

Multiple morphosequences can be recognized in a single valley system by mapping a shingled pattern of graded surfaces (Fig. 2). In general, the older sequences will have a higher elevation and may be the least preserved due to downcutting adjustments made by the next younger sequence. In many areas, such as the Connecticut River Valley, proglacial lakes provided the base level control for both ice contact and non-ice contact sequences, alike. Elsewhere, temporary dams of either till, glacial fluvial sediment, or more slowly decaying ice blocks inhibited drainage away from the ice.

The volume of glacial fluvial and lacustrine deposits found choking the valleys of western Massachusetts is strong evidence that meltwater played a major role in the deglaciation of New England (Gustvason and Boothroyd, 1987). The nature of retreat from place to place can be inferred by the types of depositional environments seen graded to a single head of outwash. In the Connecticut River Valley, kame terraces constructed between the valley wall and the ice commonly grade to ice-contact (kame) deltas built directly into Glacial Lake Hitchcock (Gustvason et al., 1975, Fig. 3). Elsewhere, broad braided meltwater streams supplied sediment to non-ice contact deltas. Associated with this deposition was the development of eskers in the stagnation zone that acted as major conduits for channelling meltwater from within and under the Ice Sheet. In the lacustrine environment, fine-grained material was carried out from the proximal delta front by overflows or interflows to be deposited as rhythmites in the distal portions of the lake (Ashley, 1972).

On this field trip we will examine a series of landforms that typify the style of deglaciation in this portion of the Connecticut Valley (Fig. 4, stop outline). The interpretation of many of these landforms is based entirely upon gross lithologic character and topographic position. Although not ideal, sedimentological aspects can be inferred from well-studied landforms elsewhere in the valley.

The following topographic maps (7 1/2 min Quads; Scale 1:25,000) are useful for self-guided tours of this trip. From north to south and east to west they include:

<table>
<thead>
<tr>
<th>North</th>
<th>Northfield</th>
<th>East</th>
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</thead>
<tbody>
<tr>
<td>West</td>
<td>Bernardston</td>
<td>Millers Falls</td>
</tr>
<tr>
<td>Williamsburg</td>
<td>Greenfield</td>
<td>Shutesbury</td>
</tr>
<tr>
<td>Easthampton</td>
<td>Mt. Toby</td>
<td>Belchertown</td>
</tr>
<tr>
<td>South</td>
<td>Mt Holyoke</td>
<td></td>
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</tbody>
</table>
Diagrammatic profile of a fluvial-lacustrine ice-contact sequence (type FLC).

Diagrammatic profile of a fluvial-lacustrine non-ice-contact sequence (type FLNC).

Diagrammatic profile of a lacustrine-fluvial ice-contact sequence (type LFC).

Diagrammatic profile of a fluvial ice-contact sequence (type FC). Detached ice blocks represent sites of future ice-contact slopes.

Diagrammatic profile of a fluvial non-ice-contact sequence (type FNC).

Diagrammatic profile of a lacustrine ice-contact sequence (type LC).

Fig. 1. Illustrations outlining the eight basic types of morphosequences as originally defined by Koteff (1974).
Fig. 2. The shingling effect of nesting morphosequences is well illustrated by this cross section by Koteff and Volckmann (1974) from the Pepperell Quadrangle, MA. The steep upice margin of each morpho-sequence represents a temporary head of outwash.

Fig. 3. Schematic sketch of Glacial Lake Hitchcock reproduced from Ashley (1972). Ice contact and non-ice contact deltas graded to the Lake provide critical evidence for understanding the deglaciation of the Connecticut River Valley.
Fig. 4. Regional map showing the location of stops outlined in this section of the guidebook.
FIELD TRIP STOPS

STOP 1. NORTH VALLEY ROAD KAME TERRACES. Pelham, MA, Shutesbury Quadrangle. From Amherst town common, proceed east 2.6 miles on Main Street toward Pelham. Turn left (northeast) onto North Valley Rd. and proceed 2.0 miles to open sand pit on the right (fig.5).

Deglaciation of the Pelham Hills probably preceded retreat of the main Connecticut Valley ice lobe. Throughout the Pelham Hills pockets of stratified sand and gravel form a complex series of kame terraces marking the retreating position of the ice. Detailed studies of the sedimentology of the terraces have not been done but reconnaissance work by Tom Rice suggests that many of the kames carried downhill into or under the ice to join englacial or subglacial meltwater channels. Flow probably consisted of a combination of lateral meltwater flow and meteoric water draining the exposed uplands. At the time these kame terraces were being deposited, ice in the main valley probably extended as far south as the Holyoke Range and flowed south on the west side of the Mt. Tom Range (cf., Larsen and Bartshorn, 1982, his fig.7).

Fig. 5. Topographic map of a small area in the Pelham Hills where numerous ice-contact glacial fluvial terraces occur in a semi-parallel series.
STOP 2. AMHERST DUMP DELTA, Amherst, MA, Belchertown Quadrangle. From North Valley Rd. turn right (west) and proceed 0.3 miles and turn left (south) onto Harkness Rd. At the end of Harkness Rd. (2.2 miles) turn right (northwest) onto Rte. 9. Proceed 0.4 miles to dump entrance on the right (north side of road) (fig. 6). Note: Dump open Mon-Sat. 7:30-4:30; closed on Sundays.

While enroute to stop 2 via Harkness Rd., note the relatively flat topography of the area on either side of Harkness Rd. Much like a braided meltwater stream, this route takes you along the surface of a kame terrace pockmarked by numerous kettle holes. These kettles now form small lakes that add atmosphere to the housing subdivision built engulfing the area.

Numerous kame terraces in the Connecticut River valley terminate in deltas graded to temporary bodies of standing water or to the stable phase of Glacial Lake Hitchcock. Similar glacial fluvial systems have been observed adjacent to modern glaciers (fig. 7). In general, the unconformable contact between topset beds (emergent glacial fluvial plain) and foreset beds (submergent prodelta slope) is used as the best estimate of the lake level into which the delta was built.

The topset/foreset contact (hereafter, T/F contact) exposed in the Amherst dump is typical of a complex suite of delta lobes clustered against the Pelham Hills east and north of the Central Vermont railroad (fig. 6). Caggiano (1977) mapped the delta complex during his work in the Belchertown Quadrangle. He interpreted the sequence as a series of delta lobes whose point of entry shifted farther north as the ice retreated. Each major point of meltwater discharge into the ice marginal lake is, thus, recorded by the axis of each lobe (Caggiano, 1977, pg. 96). Only the northwestern lobes of the delta are graded to the stable phase of Lake Hitchcock at an altitude of about 280 ft. (84.78 m asl) in this area. In contrast, the delta lobes beneath the North St/Warren Road intersection and the Dwight Cemetery are nearly 70 ft (21.2 m) higher. These older delta lobes must be graded to a small body of water confined by ice flowing through the narrow gap between the Pelham Hills and the east end of the Holyoke Range. According to the morphosequence concept of Jahns (1953) and Koteff (1974), such a chronology of meltwater drainage can be determined by the decreasing elevation of adjacent kame terraces or changes in drainage outlet.
Fig. 6. Map showing the spatial extent of an ice-contact delta complex southwest of Amherst.
Fig. 7. By analogy with kame terraces and kame deltas draining the Crillon Glacier, Alaska, the kame complex shown in Figure 6 is thought to have developed from east to west as the ice front retreated northward. (From Gaggiano, 1977).
STOP 3. **MT. WARNER DELTA.** Town of Hadley, MA, Mt. Holyoke and Mt. Toby quadrangles. From the Amherst Dump, travel west/northwest on Rte. 9 4.5 miles. Turn right (north) onto North Maple Road. Proceed 1.0 mile and turn left (west) onto Huntington Rd. Turn right (north) onto Breckenridge Rd. Near the top of the hill at 0.1 miles turn left into a forked gravel road. The right fork will take you to the top of the delta; the left fork will take you into the pit of the delta (fig. 8).

The Mt. Warner Delta represents an ice-contact delta situated in the middle of the Connecticut River Valley built directly into glacial Lake Hitchcock by meltwater from an englacial or subglacial channel system (fig. 9). The deposit averages 76 m in thickness spread over an area of about 2 square miles (5 square km). McIlvride (1982) described the Mt. Warner Delta as a part of his Masters Thesis on the northeastern part of the Mt. Holyoke Quadrangle.

The foreset beds in the delta are largely composed of reddish-brown coarse to very coarse sand, with lesser amounts of pebble gravel and fine sand. Dip angles on the beds trend westward and southwestward at angles of 24-29 degrees through at least 15 m of exposed section. Topset beds in the section are composed of yellowish-brown to reddish-brown sand and pebble gravel that thickens to the northeast from 0.5 meters at the outermost topsets to 2.0 m in the center of the delta. The bedding in the topsets varies from poorly to well-developed with interstratified sands and gravels and channel scour-and-fill structures. Interbedded with the topsets are sediments of debatable origin thought to represent flow till.

The T/F contact in the Mt. Warner Delta occurs at 278 ft. (84.7 m) based upon measurements by Koteff and Larson (see Koteff et al. figure 4, page 177, this volume. This elevation falls on a line within the range representing the stable phase of Glacial Lake Hitchcock.

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Fig. 8. Map of the Mt. Warner delta showing the ice-contact slope on the north-northeast side of the deposit.
Fig. 9. Sketch showing the relationship between topset, foreset, and bottomset beds in a typical glacial lake delta (from Ashley, 1985).
For many years two physically distinct tills have been recognized throughout much of New England (Koteff and Pessl, 1985). The "lower till" is characterized by a higher clay content (15-17%), and olive brown color, whereas the "upper till" is characterized by a very low clay content (about 4%) and a more olive gray color (Table 1, from Newton, 1978). The origin and age of these two tills has been the fuel for much debate. The more traditional view held by Schafer and Hartshorn (1965), Stone (1974), Thompson (1975), Newton (1978), and Koteff and Pessl (1985) maintains that each till represents a distinct glacial episode in New England. The alternative view is that the upper coarse grained sandy till represents an ablation phase and the lower clay-rich till a lodgement phase of the same glaciation (Denney, 1958; Goldthwait (1968), Drake (1971). Although we will not debate the issues here, let it be known that in the minds of many local workers there is still a "two-till problem" (Schafer and Hartshorn, 1965; Koteff and Pessl, 1985).

Newton (1978) carried out some of the most quantitative work on the till stratigraphy of Massachusetts and surrounding terrains. Throughout western Massachusetts, he redefined the lower till as the Thomaston till, comprised of an oxidized and unoxidized facies of similar lithologic character (fig. 11). The upper till, in contrast, was defined as the Bakersville till which he subdivided into three subfacies on the basis of compactness, texture, and the presence of fluvial bedding. Subdivision of the two till sequence was further supported by differences in bulk chemistry arising from a contrast in mineralogy.

The section at Stop 4 exposes the oxidized and unoxidized facies of the "lower" Thomaston till. What is unusual about this locality is that the unoxidized facies overlies the oxidized facies, the exact reverse of what is seen at most other sites. Although glaciotectonic thrusting is one possible explanation for the inversion of stratigraphy (fig. 12), the chemistry, mineralogy, and petrographic data of the sediments suggests that the entire section has been, in fact, overturned. Fig. 13a shows the normal trend in the depletion of base cations through an undisturbed profile; Fig. 13b, however, illustrates the observed trend in potassium and aluminum with the inverted profile. Had the section experienced overthrusting, the expected trend in base cations would be as shown in Fig. 13c. Clastic dikes throughout the section suggest that overturning, perhaps on the axis of a recumbant fold, was also accompanied by overpressuring.
Fig. 10. Location of the Leeds till site described by Newton (1978).
<table>
<thead>
<tr>
<th>CHARACTERISTIC</th>
<th>UPPER TILL</th>
<th>LOWER TILL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Texture of matrix</td>
<td>66 percent sand</td>
<td>58 percent sand</td>
</tr>
<tr>
<td></td>
<td>30 percent silt</td>
<td>27 percent silt</td>
</tr>
<tr>
<td></td>
<td>4 percent clay</td>
<td>15 percent clay</td>
</tr>
<tr>
<td>Color</td>
<td>light olive gray to olive (5Y 6/1-5/3)</td>
<td>olive brown to very dark gray (5Y 4/3-3/1)</td>
</tr>
<tr>
<td>Compaction</td>
<td>slight to moderate</td>
<td>moderate to extreme</td>
</tr>
<tr>
<td>Layering</td>
<td>horizontal textural layering common</td>
<td>occasional horizontal fine-grained textural layering. Also vertical sand dikes</td>
</tr>
<tr>
<td>Joints</td>
<td>generally not present</td>
<td>common subhorizontal jointing closely spaced near surface producing fissility. Widely spaced subvertical joints also occur</td>
</tr>
<tr>
<td>Macrofabrics</td>
<td>northeast to northwest</td>
<td>north to northwest</td>
</tr>
</tbody>
</table>
Fig. 11. Composite section of the regional till stratigraphy as interpreted by Newton (1978).
Fig. 12. In other areas of New England, overthrusting is commonly relied upon to explain inversions in the till stratigraphy. At Stop 4, chemical, mineralogical, and petrographic data argue against this explanation, in favor of overturning the entire section as if on the limb of a recumbant fold (Newton, 1978).
Fig. 13. a) Vertical changes in the $K_2O/Al_2O_3$ ratio of the $<1$ micrometer fraction of tills from the type Thomaston dam site (Newton, 1978).

13. b) Variation in the proportion of base cations that would be expected if the unoxidized zone had been overthrusted onto the oxidized zone (from Newton, 1978).

13. c) Observed variation in base cations with depth from the Leeds site. The trend shown is exactly opposite to that observed in other sections of the Thomaston till (Newton, 1978).
STOP 5. LAKE HITCHCOCK VARVES. Sunderland, MA, Mt. Toby Quadrangle. Return to Northampton interchange with I-91 and proceed north. Exit off interchange 24 to Rte 116 East. Before crossing the Sunderland Bridge over the Connecticut River at 1.3 miles, turn left (north) onto River Rd. Follow this road north 1.8 miles to a small stream draining a fenced pasture (fig. 14).

The floor of the Connecticut River Valley is lined with glaciolacustrine rhythmites. These deposits were first studied in intimate detail by Antevs (1922) who strove to duplicate the detailed Holocene varve chronology erected by De Geer (1912) in parts of northwestern Europe. Since that time, papers by Ashley (1972, 1975) have become the classic references for outlining the sedimentology of the varves and how they vary spatially throughout the Valley.

Ashley (1975) distinguishes types of rhythmic bedding in Lake Hitchcock and other glacial lakes on the basis of the relative thickness of fine versus coarse layering. With increasing distance from a point of sediment influx, rhythmites become increasingly thinner and are more strongly dominated by clay and silt deposition. In summer, coarser layers may be deposited from sediment underflows whereas silt and clay may be deposited throughout the lake and over bathymetric highs by overflows and interflows. During the winter when lakes are generally frozen over, a thin fining upward clay layer evenly blankets the more variable summer layer throughout the lake. In general, sites most distal from ice-contact and meltwater deltas in the Valley show only minimal influence from density driven underflows. During the life of Lake Hitchcock, the sediment exposed at Stop 5 was far beyond the direct influence of most sediment underflows, hence the coarser sand layers are thin.
Fig. 14. Location of stop 5 where lacustrine clays of Lake Hitchcock are exposed by a small stream eroding the valley slopes.
STOP 6. CRANBERRY POND. Mt. Toby State Forest Reserve, Mt Toby and Greenfield Quadrangles. From Stop 5, cross the Sunderland Bridge over the Connecticut River on Rte. 116. Turn left (north) onto Rte. 47 and proceed 3.9 miles to Reservation Rd. on the right. Follow Reservation Rd. 0.8 miles and park at Cranberry Pond (fig. 15).

The morphosequence concept of mapping ice retreat is best illustrated at this site where a well-developed, sharp crested esker terminates at a head of outwash. This braided outwash surface ultimately drained south through Long Plain Valley to terminate in a non-ice contact delta graded to Glacial Lake Hitchcock. Through a series of cartoons (drafted by DUW) we demonstrate our interpretation of the morphostratigraphic relationships observed in this valley.

1. A preglacial view across the Connecticut River Valley may have looked like this with the proto-Deerfield River draining directly into the Connecticut south of Sugarloaf Mt.
Fig. 15. Location of Cranberry Pond and associated esker, kame, and outwash surfaces.
2. During the period from roughly 25 ka to 16 ka BP glacial ice of the Connecticut Valley lobe overwhelmed the landscape. Ice retreat from Long Island Sound may have begun as early as 19,500 years.

3. By ca. 15,000 years BP glacial ice began to retreat from the Mt. Toby area. Eventually the ice front was subdivided by Mt. Toby into one large lobe in the main valley and a small lobe pinned between Mt. Toby and the adjacent highlands. Contemporaneous with this retreat was the incipient development of a broad braided outwash plain that terminated in Lake Hitchcock.
4. Within a few tens of years, a large ice mass stagnated east of Mt. Toby and gradually became buried by outwash. At this same time the Sunderland delta continually prograded south into the Lake and the Potumtuck Range emerged from beneath the main ice lobe, splitting the ice front into two lobes.

5. By 12.5 ka to 13 ka, the Laurentide Ice Sheet had retreated into the St. Lawrence Lowlands far to the north. At about this same time, the drift dam at Rocky Hill, Conn., eroded and Lake Hitchcock drained. The Long Plain Brook quickly eroded through the Sunderland delta producing a small fan at the foot of the prodelta slope. Because glacial fluvial sediments now clog the valley west of Mt Sugarloaf, the Deerfield River was diverted northward around the Potumtuck Range.
STOP 7. SUNDERLAND DELTA. Sunderland MA, Mt. Toby Quadrangle. From Cranberry Pond proceed 5.1 miles south on Rte. 63 and turn left (west) onto a dirt road opposite Juggle Meadow Rd. Go passed the farmhouse on the right and proceed 0.2 miles to park next to two tobacco barns (fig.16). Note: there are four barns shown on the map.

The Sunderland Delta (or Long Plain Delta) is a non ice-contact delta fed by meltwater draining the ice east and north of Cranberry Pond. The T/F contact here has an elevation of 295 ft. (89.9 m asl). With retreat of the ice, the delta prograded into the lake (fig. 17) and the esker north of Cranberry Pond was gradually unroofed.

Following drainage of Glacial Lake Hitchcock at Rocky Hill, the Long Plain Brook quickly adjusted to a lower base level by incising a small canyon through the delta. This downcutting produced an alluvial fan out over the floor of the former lake (fig.18). Today ground water maintains a continuous flow through the delta, producing natural springs from sand and gravel beds confined by lake clays at the foot of the delta slope (fig. 19). For this reason, the State maintains a Fish Hatchery for trout in this area of the valley.

STOP 8. BARTON COVE PLUNGE POOLS. Turners Falls, MA, Greenfield Quadrangle. Follow Rte. 63 north 11.4 miles to Rte.2 through Millers Falls. Proceed west on Rte. 2 3.0 miles and turn left (south) into Barton Cove Park (fig. 20).

Ice retreat up the Connecticut River Valley north of Turners Falls produced a large non ice-contact delta (the Montague Plain) and smaller associated deltas at present elevations between 300 and 350 ft. (91 and 106 m). Jahns (1966) reportedly found gravelly beach and bar deposits marking the edge of Lake Hitchcock at the fringes of several deltas where they occur below low wave-cut scarps.

The drainage of Lake Hitchcock eventually forced the post-lake Connecticut River to erode through all of the glacial fluvial and glacial lacustrine sediments that choked its former course. Due to the volume of sediment in the Montague Plain, the Connecticut initially detoured its way around the north side of Canada Hill and drained via Greenfield before establishing its present course (Jahns, 1966). As it entrenched, the river gradually became superimposed onto the bedrock peninsula of Triassic age Turners Falls Sandstone that now protects Barton cove. Plunge pools on the northwest side of the peninsula are interpreted as erosional scars of a waterfall that cascaded over the bedrock obstruction until an easier route was cut through the delta complex. Lower terrace surfaces surrounding the Montague Plain are thought to mark lowering levels of Lake Hitchcock.
Fig. 16. Sunderland delta of stop 7.
Fig. 17. Schematic cross-section of the evolution of Sunderland Delta, Cranberry Pond, and associated esker system. 1) Ice front at about 15,000 years BP.; 2) ice front and prograding delta tens of years later; 3) tens of years later, ice stagnates froming Cranberry Pond, the esker is exposed, and ice continues to retreat; 4) The Long Plain Brook today.
Fig. 18. The Sunderland Delta as it appears today after the drainage of Lake Hitchcock and the downcutting of Long Plain Brook.

Fig. 19. Schematic diagram showing the high water table that produces active artesian springs at the base of the prodelta slope.
Fig. 20. Map showing the spatial relationship between the plunge pools and the Montaque Plain.
STOP 9. BERNARDSTON DRUMLIN FIELD. Bernardston, MA, Bernardston Quadrangle. From Bartons Cove continue west 3.0 miles on Rte 2 to Rte 5/10. Follow Rte 5/10 north 4.8 miles to Bernardston. Turn right (east) and proceed 2.6 miles on Rte 10. Pull off in excavation on the left (north) side of the road (fig. 21).

The broad conical hills on either side of Rte 10 near Bernardston represent one of the best developed drumlin fields in this portion of Massachusetts. They are generally oriented a few degrees east of north, parallel to the regional ice flow direction, and rise 50 to 200 ft. (15 to 60 m) above the adjacent glacial fluvial plain. Although exposures are limited, most seem to bare some type of bedrock core and are thickly mantled by "lower till". Exposures at the base of the drumlin at Stop 9 are new and have not been carefully studied.
Fig. 21. Bernardston drumlin field and the location of Stop 9.
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