Late Wisconsinan Glacial Deposits in the Portland - Sebago Lake - Ossipee Valley Region, Southwestern Maine

Guidebook for the
58th Field Conference of the Northeastern
Friends of the Pleistocene

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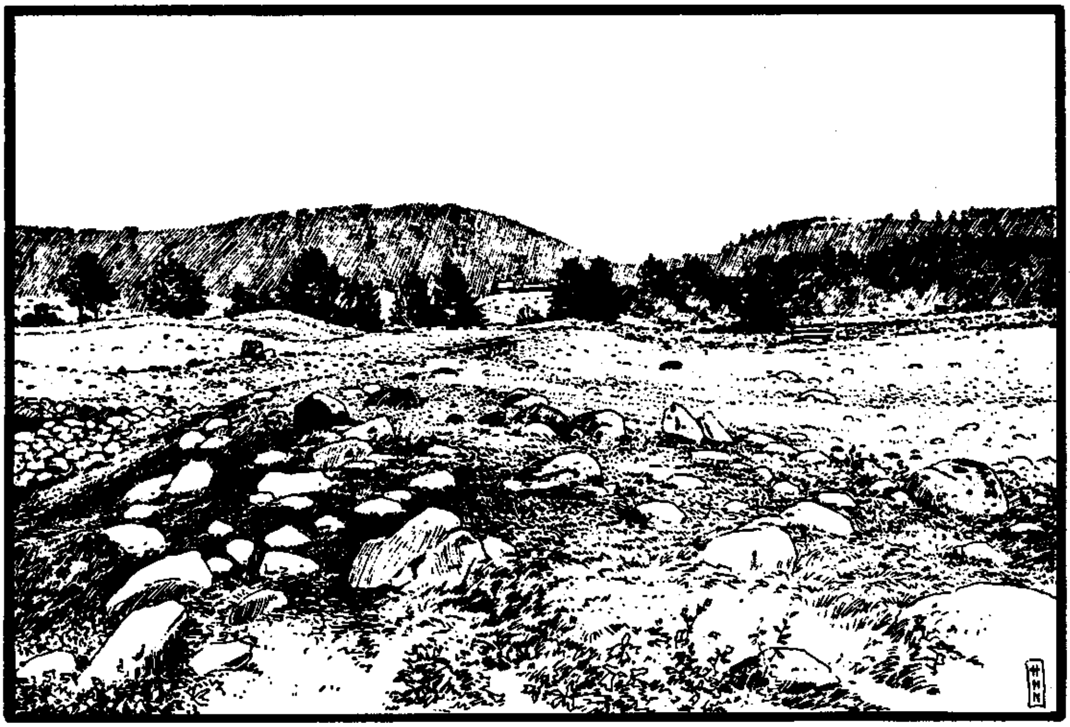

A. RETICULATED RIDGES OF COARSE WATER-ROLLED GRAVEL; PARSONSFIELD. LOOKING NORTH.
Glacial river flowed through low pass in distance down the hill to the foreground.

Woodrow B. Thompson, *Maine Geological Survey*
P. Thompson Davis, *Bentley College*
John C. Gosse, *Los Alamos National Laboratory*
Robert A. Johnston, *Maine Geological Survey*
Robert Newton, *Smith College*

1995

Cover illustration from *The Glacial Gravels of Maine*, by George H. Stone (1899). The location of this view is uncertain. It is probably in the northeast part of the West Newfield Quadrangle, showing deposits south of the gap between Cedar Mtn. and the unnamed hill to the east.

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Portland, Maine
May 12-14 , 1995

Field Trip Leaders: Woodrow B. Thompson, Maine Geological Survey
P. Thompson Davis, Bentley College
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1995

FOREWORD

The 1995 Friends of the Pleistocene field conference in Portland marks the fifth time that the Friends have gathered in Maine during the 58 years in which they have held trips. The previous Maine reunions were led by: Art Bloom (1961 - Portland region); Hal Borns (1967 - eastern coastal Maine); Woody Thompson and Jeff Smith (1983 - Augusta region); and Steve Kite and Tom Lowell (1988 - northern Maine). The Maine Geological Survey is pleased to welcome the Friends of the Pleistocene back to our state. We hope that you will find much of interest at the localities described in this guidebook. It is the outgrowth of several years of field work conducted by the trip leaders, primarily for the MGS surficial quadrangle mapping program.

Users of this guidebook who are not familiar with the Friends of the Pleistocene may wonder about the background and purpose of the Friends. The Friends are an informal organization with a mutual interest in Quaternary science (chiefly glacial geology). Every spring the Friends meet in a different area to examine and discuss field-oriented problems under the leadership of one or more hosts. Originally consisting mainly of college professors, the group of participants has grown over the years to encompass students of Quaternary geology and other persons with similar interests. As Dick Goldthwait aptly put it, a Friend of the Pleistocene is "anyone who wants to be".

George White and J. Walter Goldthwait hosted the first meeting of the original (northeastern) Friends of the Pleistocene, which was held in New Hampshire in 1934. Since that time, the Friends have met annually - except during World War II - and have traveled widely through the northeastern states and adjacent Canada. On the occasion of their 50th reunion, Dick Goldthwait published his colorful recollections of those first 50 years, including the other Friends sections that have been organized around the country (Goldthwait, 1988). An updated list of the reunions held in the Northeast is included here in the Appendix.

This guidebook has been published to provide a durable product for future reference. Much useful and interesting information from the early Friends trips has been lost due to lack of a guidebook or surviving handout. However, the authors wish to emphasize that we are reporting on work in progress. In the spirit of the Friends of the Pleistocene, we are presenting a mixture of field observations, some reasonably well-founded conclusions, and remaining problems concerning the deglaciation of southwestern Maine. Your input is encouraged at each of the stops described here, and to stimulate discussion we have posed questions for many of the stops.

Note: Nearly all of the stops in this guidebook are located on private property. Some of them are working gravel pits, and frequent visitors may not be welcome. Ownership and operational status of sites are subject to change. Permission was granted to visit them during the 1995 Friends of the Pleistocene trip, and must again be obtained by other persons or groups for subsequent visits!

ACKNOWLEDGMENTS

The authors thank the Maine Geological Survey, and especially Walter A. Anderson (State Geologist), for supporting their field work and preparation of this guidebook. Robert D. Tucker and John B. Poisson, of the Survey's Earth Resources Information Division, assisted with photo reproduction and many other details in the publication process. Helpful suggestions for conference logistics were provided by Carl Koteff, Duane Braun, and other Friends of the Pleistocene. Christopher Dorion and George Jacobson, Jr. (Institute for Quaternary Studies, University of Maine) assisted in preparing and identifying fossil plant samples for radiocarbon dating.

The surficial geologic quadrangle mapping that provided the basis for much of this trip was funded in part by the U. S. Geological Survey's Cooperative Geologic Mapping Program (COGEOMAP) and its successor, the State Geologic Mapping Program (STATEMAP). The National Science Foundation, through a grant to the University of Maine under the EPSCoR Program, has supported a joint UM-MGS project to acquire and compile data on the timing of deglaciation throughout the state. Radiocarbon dating of plant fossils from Stop 3 was carried out at the National Ocean Sciences AMS Facility in Woods Hole, Massachusetts, with support from NSF Cooperative Agreement No. OCE-801015.

We are very grateful to the companies listed below for permission to visit their sand and gravel pits, both for mapping and research purposes, and to include their properties in the 1995 Friends of the Pleistocene itinerary. Without their cooperation, and the assistance of many other landowners, our field projects would have been much more difficult.

Ronald Barrett -- Brownfield, Maine
Dearborn Brothers Construction, Inc. -- Gorham, Maine
William Guptill -- Baldwin, Maine
P & K Sand & Gravel, Inc. -- Naples, Maine
Resource Investments, Inc. -- West Lebanon, New Hampshire
C. R. Tandberg, Inc. -- North Windham, Maine
Alan and Sam Toppi -- Westbrook, Maine

We also acknowledge the staff of Sebago Lake State Park for arranging the use of their facilities.

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INTRODUCTION

The area covered in this field guide is a region of southwestern Maine extending from Portland westward to the New Hampshire border (Figure 1). It encompasses part of the coastal lowland, from Portland to Sebago Lake, and portions of the Saco and Ossipee River valleys in the hilly terrain west of Sebago Lake. The Portland 1:100,000-scale quadrangle map covers the entire area, while the Surficial Geologic Map of Maine (Thompson and Borns, 1985) provides a broad overview of the Quaternary deposits.

The trip described here examines contrasting glacial deposits that formed in glaciomarine and terrestrial environments. The authors' work in southwestern Maine will provide a framework for discussion of problems relating to the timing and modes of deglaciation. Most of the sediments seen on the trip were deposited by glacial meltwater during the recession of the late Wisconsinan ice sheet. In the coastal lowland, the retreating ice margin was in contact with the sea and deposited a variety of glaciomarine sediments. These deposits have been studied by many people during the last twenty years. The accumulated evidence shows that the marine-based ice margin was active, but details of sedimentary processes and the chronology of deglaciation are the subjects of ongoing research. The precise relationship between ice retreat and changing sea level is being studied to help provide clues to late-glacial climate change and isostatic crustal response to deglaciation.

The central part of the field trip area straddles the inland limit of late-glacial marine submergence (known as the "inland marine limit" in contrast to the "upper marine limit", which is the maximum altitude reached by the sea in any particular location; many islands formerly existed in the submerged areas). This part of Maine has proven to be one of the more difficult areas in which to determine the location of the inland marine limit. Glaciomarine deltas occur around the east, south, and west sides of Sebago Lake, but sediments that appear to be lacustrine rhythmites are seen at the north end. The lake itself is over 90 m deep in places, and the floor is locally below sea level. The itinerary for Day 1 examines the glacial deposits surrounding Sebago Lake, as well as research on the sediments beneath the lake basin and their bearing on the position of the marine limit.

The Day 2 itinerary follows the Saco and Ossipee River valleys. Here, too, the marine limit is not obvious. A succession of coarse deltaic deposits occur at similar elevations, and in the absence of known fossil localities or typical glaciomarine mud it becomes difficult to judge whether certain of these deposits are marine or lacustrine. Several localities will be examined in this transition zone.

In central to southern New England, long-lived (and lively) controversies regarding the mode of ice retreat have been addressed by the "morphosequence concept", which evolved over many years of field mapping (Koteff and Pessl, 1981). As detailed surficial mapping progresses across the interior of southern Maine, we should consider the extent to which this concept can be applied. Can we find evidence for sustained ice flow as the White Mountain foothills were deglaciated? Are there any end moraines, or meltwater deposits that define heads-of-outwash, which formed along a distinct retreating ice margin? What was the chronology of ice retreat? The glacial lake history and mode of esker deposition will also be examined in this area.

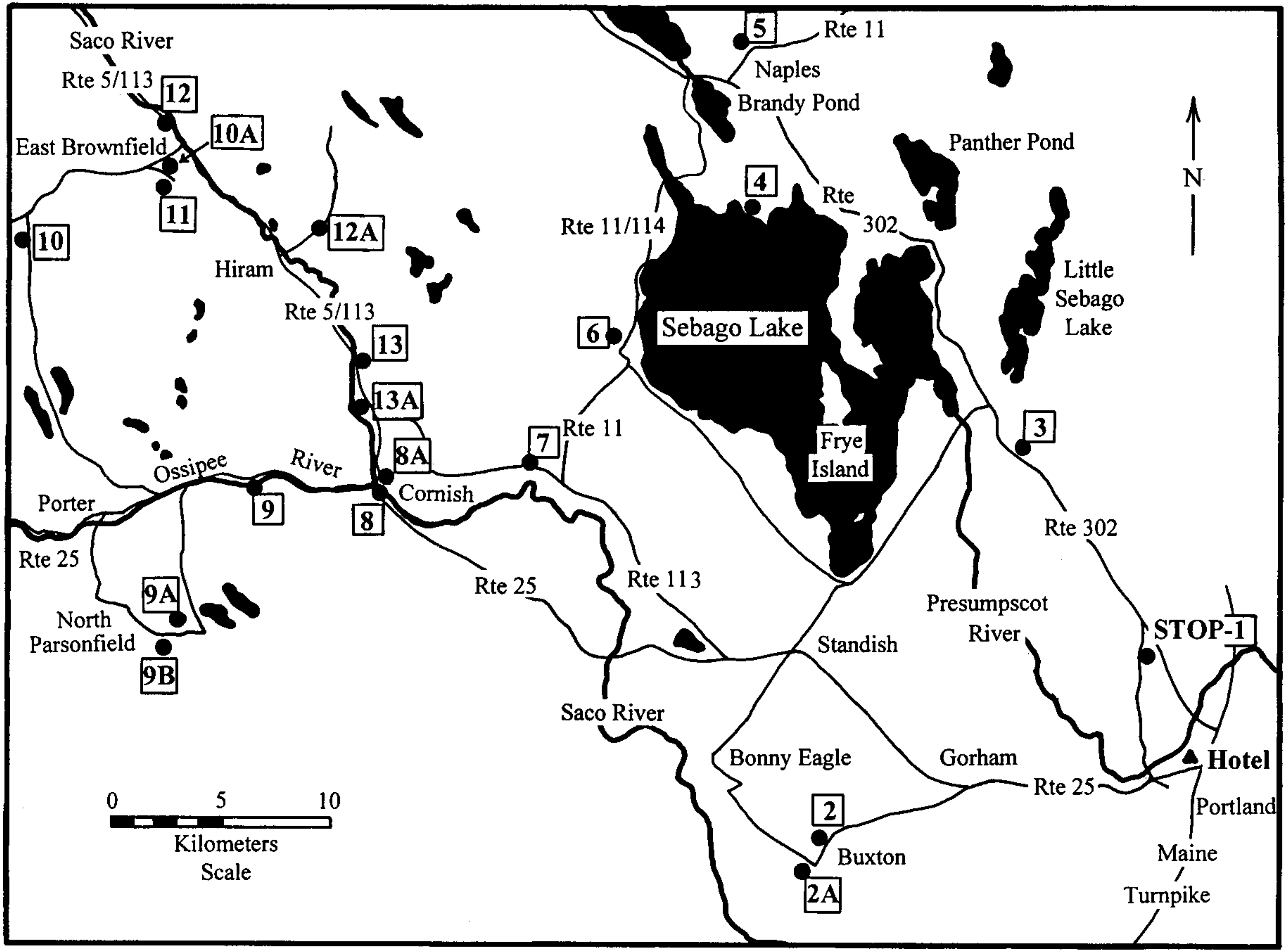


Figure 1. Map showing location of field trip stops described in this guidebook.

PREVIOUS WORK

Borns (1989) has summarized the history of Quaternary geologic investigations in Maine. The first in-depth study of Quaternary deposits in the state, including the region described in this guidebook, was conducted by G. H. Stone (1899). Stone's investigation of the "glacial gravels of Maine" was largely concerned with esker systems and related ice-contact features, but he also described the other types of glacial deposits and provided many new insights into their origin. Stone compiled the first map showing successive ice-margin positions during the deglaciation of Maine. He believed that the last ice sheet continued to flow actively during at least the early stages of thinning and recession (Stone, 1899, p. 271).

Stone often compared his models with observations of modern glacial processes, and in some cases his theories have survived the test of time better than work that followed in the early 1900's. He described the process of end-moraine deposition in the coastal lowland as resulting from stillstands of an active glacier margin, to which debris was conveyed from a thin basal zone of debris-rich ice (p. 275-276). Many of Stone's interpretations of meltwater deposits have been improved by later workers, but he correctly interpreted Sebago Lake as being dammed by a glaciomarine delta (p. 242-243). He also commented at great length on the origin of Maine eskers ("osars"), and described the esker ridges in the Brownfield area (see Stop 10, this trip).

The next study of Maine's glacial geology was published by H. W. Leavitt and E. H. Perkins in the mid 1930's. Their three-volume series combined a statewide survey of road materials with descriptions of glacial and postglacial sediments. They also compiled an accompanying map showing the surficial geology of Maine. Leavitt and Perkins (1935) recognized that numerous "moraine banks" and ice-contact "recessional deltas" indicated the progressive retreat of the ice margin during deglaciation of Maine's coastal zone (p. 194). However, they described the moraines as having been deposited by meltwater from the "stagnant marginal zone" of the glacier (p. 44). The proximal part of the glaciomarine delta at the south end of Sebago Lake was considered to be one such moraine.

Leavitt and Perkins (1935) pointed to kame terrace sequences farther inland as evidence that "In the hilly western portion of the State, the last ice remained as narrow tongues along the valley floors, and wasted away in place" (p. 193-195). The latter statement is pertinent to the heated controversy among New England glacial geologists, which peaked in the 1930's, regarding the mode of glacial retreat. Previous assumptions that an active ice margin had receded northward from New England were essentially replaced for several decades by the model of widespread simultaneous stagnation and downwastage of the ice sheet (e.g. Flint, 1930; Goldthwait, 1938).

Leavitt and Perkins (1935, p. 92-95) described meltwater deposits in the portions of the Ossipee and Saco River valleys included in our trip itinerary for Day 2. Although they advocated the stagnation model for deglaciation of this part of Maine, Leavitt and Perkins noted the preferred distribution of ice-contact glacial lake deposits in the *north*-draining tributaries of the Saco and Ossipee valleys. The succession of ice-dammed lakes in this area supports the concept of progressive, generally northward retreat of the ice-margin. Similar ponding of glacial lakes in north-draining valleys all over New England has long been used as evidence in favor of "systematic" ice retreat, especially by proponents of active-ice during deglaciation.

In the 1940's the Maine Geological Survey investigated the economic potential of glaciomarine clay deposits. As part of this project, L. Goldthwait (1949, 1951) published two reports describing the marine clays of the Portland-Sebago Lake region. He considered problems relating to the deposition of the clays, including the number of units and the timing of marine submergence relative to deglaciation. Goldthwait concluded that there is one basic clay unit, the upper part of which is oxidized to a brownish color, in contrast to previous workers who regarded the brown and blue clays as being stratigraphically distinct. Although he thought that marine transgression postdated ice retreat in this part of Maine, Goldthwait proposed that the Sebago Lake basin was still occupied by remnant ice during the submergence of areas around the lake.

A. L. Bloom (1959) wrote a popular guide to the geology of Sebago Lake State Park, in which he described late-glacial and postglacial deposits at the north end of the lake. Bloom (1960, 1963) also carried out the first detailed studies of glaciomarine sediments and glacial isostasy in southwestern Maine. He described the clusters of end moraines that occur abundantly in the zone of marine submergence, and coined the name "Presumpscot Formation" for the glaciomarine muds (Bloom, 1960). This formation name was derived from the Presumpscot River in the Portland-Westbrook area, near which Bloom found some of the best exposures. The name has been generally used by later authors for glaciomarine muds throughout southern Maine. Bloom (1963) thought that the marine transgression lagged behind deglaciation in Portland, but caught up with the receding glacier margin farther inland, where large ice-contact deltas were deposited into the sea.

Borns and Hagar (1965) and Borns (1967) concluded that the sea was in contact with the retreating late Wisconsinan ice margin in central and eastern Maine. Borns' work also demonstrated that the ice was still active as the system of end moraines was deposited in the eastern part of the coastal zone. His deglaciation model was extended to southwestern Maine during the following decade, when the Maine Geological Survey began a program of reconnaissance surficial quadrangle mapping. Field work for this program by Borns, B. G. Andersen, G. W. Smith, and W. B. Thompson, together with aquifer mapping by G. C. Prescott, Jr. (USGS), provided the basis for ongoing investigations of glaciomarine sediments in the coastal zone (e.g. Thompson, 1979). Stuiver and Borns (1975) compiled radiocarbon ages from the marine deposits to obtain an updated chronology of deglaciation and resulting crustal uplift.

Since the 1980's there has been a proliferation of research on the glaciomarine deposits in southwestern Maine. From comparison with published studies of "subaqueous outwash" in the St. Lawrence Lowland of Canada (Rust and Romanelli, 1975), it was realized that submarine fans are common in this region, and form a major component of many end moraines (Smith, 1982; Thompson, 1982). Recent work has focused on detailed facies analysis of these deposits (e.g. Retelle and Bither, 1989; Smith and Hunter, 1989). The glaciomarine deltas have also been described in more detail and yielded information on the configuration of the upper marine limit and late-glacial sea-level history of Maine (Thompson and others, 1989; Crossen, 1991; Koteff and others, 1993).

Smith (1985) analyzed the available radiocarbon ages from glaciomarine deposits, and proposed that the receding late Wisconsinan ice margin did not reach the inland marine limit in central Maine until about 12,500 yr B.P. However, dated marine shells from sediment cores indicate that the ice withdrew to Gould Pond in central Maine by about 13,300 yr B.P. (Anderson and others, 1992) and even farther north

(Mattaseunk Lake) in the Penobscot Lowland by 13,400 yr B.P. (Dorion, 1994). All ages given here are in radiocarbon years, and the marine shell ages have not been corrected for the "reservoir effect". The magnitude of this correction has yet to be determined for coastal Maine, but studies in the North Atlantic region suggest that at least 400 years should be subtracted from the marine radiocarbon ages to make them comparable with terrestrial ages (Mangerud and Gulliksen, 1975; Bard, 1988; Broecker and others, 1988).

While many new radiocarbon ages have been obtained for eastern Maine through the University of Maine - MGS EPSCoR Program, the deglaciation chronology in the western part of the state is still in an early stage of documentation. Until recently, the oldest limiting age on glacial recession in southwestern Maine was 13,830 +/- 100 yr B.P. (QL-192), from shells collected by J. T. Andrews at a coastal bluff in Kennebunk (Smith, 1985). Ages on *Portlandia arctica* shells now indicate ice retreat as far north as Scarborough by 14,800 yr B.P. (H. W. Borns, Jr., pers. comm.), and Freeport by 14,000 yr B.P. (Weddle and others, 1993).

It will be noted that much of the previous work in the field trip area has been concentrated in the Portland-Sebago Lake region. Following the Leavitt and Perkins (1935) survey, no investigations of glacial deposits inland from the marine limit were undertaken for the next 40 years. In the 1970's, G. C. Prescott, Jr. prepared surficial aquifer maps that included reconnaissance mapping of deposits in the region extending west to the New Hampshire border (Prescott, 1979, 1980). W. R. Holland updated the aquifer mapping along the portions of the Saco and Ossipee Valleys visited on this trip (Holland and Thompson, 1987), and he began detailed surficial mapping of the Brownfield, Cornish, Hiram, and Kezar Falls quadrangles. Holland was greatly interested in the glacial history of the area, and prior to his untimely death had started to investigate many of the problems discussed here (Holland, 1986). In recent years, the quadrangle mapping was resumed by P. T. Davis, R. Newton, and W. B. Thompson.

SUMMARY OF GLACIAL HISTORY

This section summarizes the glacial history of the field trip area and describes some of the types of deposits that will be examined. Detailed discussion of field localities and ongoing research problems is deferred to the individual stop descriptions. The topics discussed in this guidebook are chiefly concerned with the mode and chronology of recession of the late Wisconsinan ice sheet, and the marine and terrestrial sediments deposited during deglaciation. The late-glacial deposits are the best-preserved part of the Pleistocene record in southwestern Maine. They provide clues to the ice-sheet dynamics, isostatic uplift, and climate and sea-level change during this time period.

Most of the till deposits exposed in the field trip area are likewise products of late Wisconsinan ice. However, sites elsewhere in southern Maine show a deeply oxidized "lower till" believed to have resulted from an earlier glaciation (Weddle and others, 1989), and a few sections show what is probably this older till beneath the fresh late Wisconsinan "upper till" (e.g. Thompson, 1986; Thompson and Smith, 1988). Weddle and others (1989) provided a thorough review and discussion of the two principal tills found across much of New England, including criteria for their recognition. These authors supported an Illinoian age for the lower till, based on correlation with the section at Sankaty Head on Nantucket Island, Massachusetts (Oldale and others, 1982). Oldale and Colman (1992) reaffirmed this correlation and assigned the lower till to oxygen-isotope stage 6.

There is little stratigraphic record of climatic events in southern Maine during the Sangamonian interglacial stage or the early to middle parts of the Wisconsinan Stage. At present, there is no firm evidence that this region was glaciated in early to mid Wisconsinan time (Oldale and Colman, 1992; Weddle, 1992). Radiocarbon ages (Stone and Borns, 1986; Anderson and others, 1988; C. C. Dorion, pers. comm.) suggest that the leading edge of the Laurentide Ice Sheet was advancing across southern Maine approximately 25,000 years ago. Striations and glacially streamlined hills indicate that the dominant ice flow direction was southeast to south-southeast (Thompson and Borns, 1985a). This late Wisconsinan glaciation removed most deposits from earlier glacial episodes, except for scattered remnants of the older till and a few saprolites that survived in topographically protected areas.

As noted above, the subsequent recession of the last ice sheet caused the glacier margin to reach southwestern Maine by 15,000-14,000 years ago. Although eustatic sea level was over 100 m lower at this time (Fairbanks, 1989), isostatic depression of the region resulted in marine submergence of lowland areas as the ice withdrew. Subsurface data from ice-contact glaciomarine deltas indicate that they were deposited in shallow water -- less than 100 m and often only 20-40 m deep -- and thus formed along a grounded tidewater-glacier margin (Thompson and others, 1989). The orientations of end moraines and other ice-marginal deposits show that the direction of glacial retreat was north to north-northwest. Exposures in end moraines often reveal ice-shove structures and lodgement till with erosional basal contacts, demonstrating that the ice remained active in the marine environment (Smith and Hunter, 1989).

Hundreds of end moraines were deposited in the Portland-Sebago Lake area. Especially fine examples are seen in Buxton, Gorham, and Windham (e.g. Stop 2-A). These moraines commonly occur as clusters of parallel ridges. Many of them are more-or-less concealed beneath younger ice-marginal deposits (fans and deltas) or a veneer of glaciomarine mud (Presumpscot Formation). The moraine ridges in coastal Maine are generally 1-15 m high. They range in width from 5-15 m to over 100 m in the large stratified moraines. Lengths range from 50-100 m to several kilometers. Many of the examples near Portland are minor (DeGeer) moraines, which are typically just a few meters high and up to several hundred meters long. The moraines show considerable variability in composition and depositional process. The smaller ones may consist entirely of glacial diamict, including till facies deposited by lodgement and subaqueous debris flows, but many moraines contain sand and gravel deposited as submarine fans (Stop 3).

Larger individual submarine fans are also common in the zone of marine submergence (Stop 2). Diamict lenses and coarse gravels occur in the ice-contact parts of these deposits, which formed at the mouths of glacial ice tunnels. Distal facies usually consist of well-stratified sand interbedded with Presumpscot Formation muds. Fan deposition occurred through several processes, including traction currents of sediment-laden meltwater discharged from the tunnel mouths (density underflows), slump-generated turbidity currents, and settling of fine sediments from interflow or overflow plumes. (Retelle and Bither, 1989).

Glaciomarine deltas formed in the shallow ocean waters bordering the Sebago Lake basin (Stop 3), and in the Saco River valley. They are coarse-grained Gilbert-type deltas, with fluvial topset beds (delta-plain deposits) overlying foreset beds deposited on the prograding delta front. Many of the deltas in Windham and around the southern part of Sebago Lake were formed at the ice margin, often where it was pinned against bedrock strike ridges or other topographic highs. These ice-contact deltas probably received most

of their sediment from ice tunnels. Eskers connect with the proximal margins in some cases. The surveyed elevations of the topset/foreset contacts define the upper marine limit, which is at a present elevation of about 90-95 m (higher to the northwest) in the Sebago Lake area (Thompson and others, 1989; Thompson and Koteff, unpub. data).

As the glacier margin continued to retreat, a different assortment of meltwater deposits formed inland from the marine limit (area covered in Day 2 itinerary). They include esker systems (Stop 10) and other glaciofluvial deposits, as well as glaciolacustrine deltas (Stops 10-A, 11, 13) and fine-grained lake-bottom sediments (Stop 12). Lakes were impounded in north-sloping ice-dammed valleys, and upstream from temporary drift dams in the Saco Valley. The Ossipee and Saco River basins locally contain well-developed stratified-drift morphosequences, whose implications for the style and pattern of deglaciation are discussed in this guidebook.

Definite end moraines are not common in the area above the marine limit, but some probable examples have been identified in the Hiram Quadrangle. One of the larger moraine ridges is seen at Stop 12A. Several valleys contain numerous short, stubby ridges of glacial diamicton and/or sand and gravel. The origin and spatial relationship of these ridges to the contemporaneous glacier margin are not obvious. Holland (1986) classified them into two types: *ribbed moraine*, consisting mainly of diamicton ridges oriented perpendicular to the ice-flow direction, and *ice-disintegration moraine*, consisting of hummocky deposits of bouldery diamicton, sand, and gravel. Examples of these deposits are seen at stops 9A and 9B, located in the Kezar Falls Quadrangle.

Glacial meltwater channels are commonly found in the hilly terrain of interior southwestern Maine. Three types of channels occur in this area: (1) Proglacial channels formed where meltwater streams drained through cols in the hills adjacent to the ice margin, or incised earlier meltwater deposits in valleys. These channels sometimes became later spillways for lakes that were dammed on their proximal sides as the ice continued to retreat. (2) Lateral channels were cut into till deposits on hillsides adjacent to valley ice tongues. They slope obliquely along the hillside with profiles that probably conform to the former gradient of the ice margin, and often occur in groups that reflect the thinning and retreat of the ice. (3) Engorged channels formed where meltwater streams plunged down under the ice. They may trend directly downslope and typically are incised in glacial sediments on the lee sides of hills.

In contrast to the coastal lowland, we have no radiocarbon ages that closely limit the time of deglaciation in the area visited on Day 2. This situation is expected to change in the near future, when lake-bottom sediments are cored and dated. Considering the marine shell ages mentioned above, we infer that this area was uncovered sometime between 14,000 and 13,000 yr B.P. The lake that formed during ice retreat from the Saco Valley upstream from Hiram (Lake Pigwacket) persisted for considerable time after deglaciation. Plant remains from three test borings in Lake Pigwacket sediments at Fryeburg yielded AMS radiocarbon ages of 11,680 +/- 105 to 11,255 +/- 80 yr B.P. (AA-4769, AA-10157, and AA-10158; Thompson, in prep.) Hu (1989) identified a variety of plant pollen and macrofossils in samples from these borings. He concluded that they came from a forest environment indicative of a warm, nonglacial climate. Hu's qualitative analysis suggests that the plant material dates to the Allerod warm interval, immediately preceding the Younger Dryas.

ROAD LOG FOR DAY 1

Portland - Sebago Lake Area

Mileage (cumulative and incremental)

Cum	Inc	
0.00	0.00	Depart toll booth at Exit 8 on Maine Turnpike in Portland. Go straight ahead on Larrabee Rd., which curves to L and quickly meets Business Rte. 25.
0.40	0.40	Turn R onto Business Rte. 25 and go W into town of Westbrook.
1.10	0.70	Keep R at yellow blinker.
2.00	0.90	Turn R at light, onto Bridge St. (= Methodist Rd.) and proceed N.
5.30	3.30	Turn R into the Toppi Pit.

STOP 1: TOPPI PIT - Striated bedrock, end moraines, and glaciomarine sediments (Portland West Quadrangle)

Leader: Woodrow Thompson

This stop is located in Westbrook, in the Mill Brook valley between Methodist Road and U.S. Route 302. The north-south valley has been filled with glaciomarine sediments, while till is exposed on the higher ground on either side. The elevation of the upper marine limit in this area during deglaciation was approximately 285 ft (87 m) (Thompson and others, 1989). Numerous end moraines occur just north and west of the Toppi Pit (Figure 2). Exposures at this stop and other nearby pits indicate that many other moraines are buried beneath the marine sediments along Mill Brook. The dissected upper surface of this valley-fill (the former sea floor) can be seen along the power line as we drive into the Toppi Pit. The features to be examined at this stop include (1) a large expanse of glacially abraded bedrock, and (2) a sequence of glacial and glaciomarine deposits that record the recession of the marine-based late Wisconsinan ice margin from the Mill Brook Valley.

Five major units (including bedrock) are present in the Toppi Pit. Excellent sections were exposed in late 1992, when the site was documented for a Geological Society of America field trip (Weddle and others, 1993). At that time, the eastern part of the pit was being operated on four levels, each exposing 3-5 m of section. The working faces trended north to east-northeast across the pit, and showed oblique cross-sections through two end moraines. These moraines were situated on the striated bedrock surface, and were overlain by submarine fan deposits and glaciomarine mud. The exposures have changed greatly during the last two years. Deposits seen during the 1993 GSA trip have been largely removed, while new sections have been opened in both the northern and southern parts of the pit area (see update below). The basic stratigraphy is unchanged, and is typical of end moraine/submarine fan associations formed during the deglaciation of coastal Maine. The units in the Toppi Pit are described below, from oldest to youngest, as they were exposed in 1992-93.

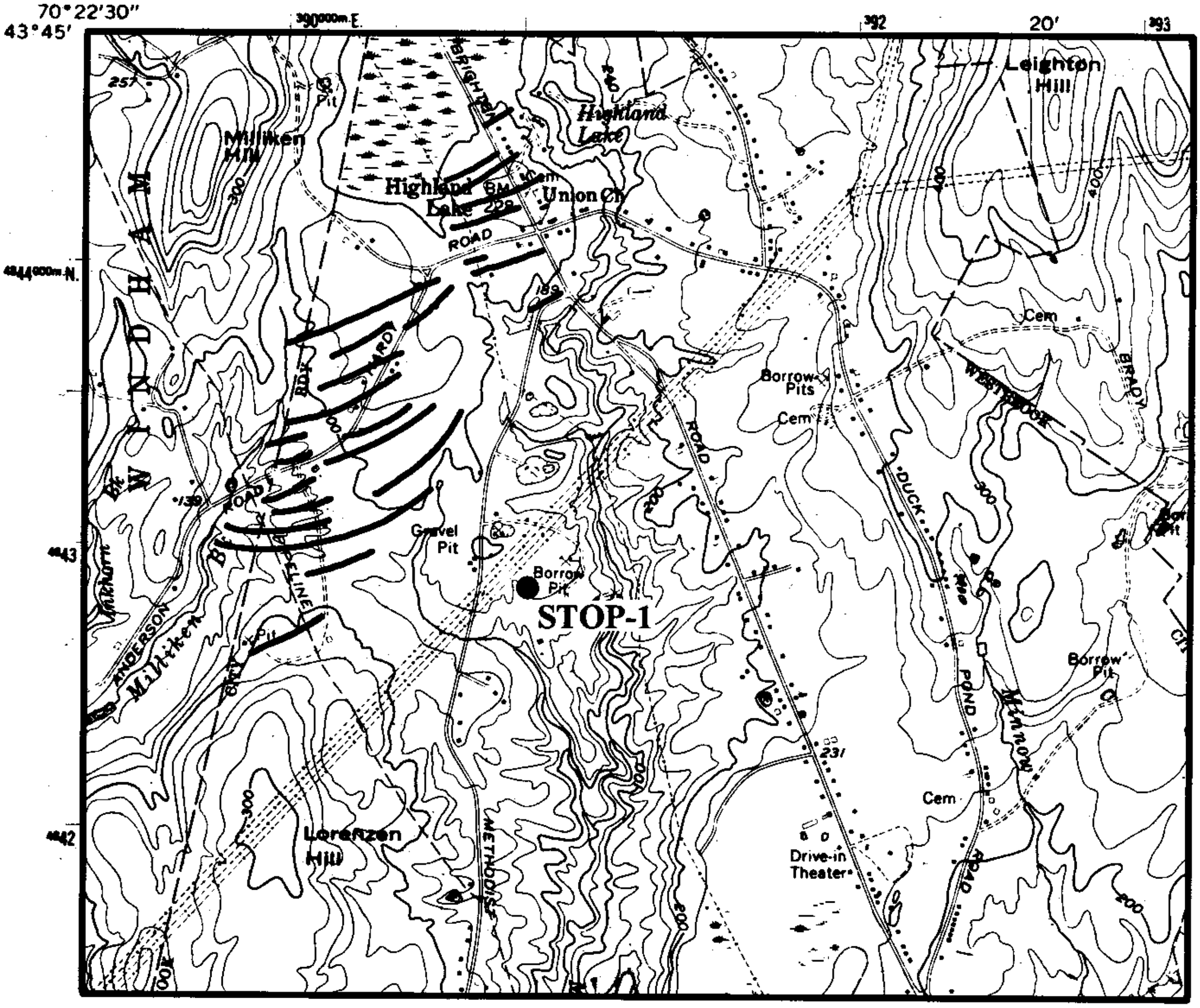


Figure 2. Location of Stop 1 (Toppi Pit, Westbrook). Lines indicate crests of end moraines that are not buried by glaciomarine sediments.

Unit 1 - bedrock. The bedrock in the pit floor is chiefly granite and granite pegmatite. The outcrops consist of a series of rock knobs, on which the stoss sides are prominently striated and polished, while the lee sides commonly show a smooth, scalloped surface that presumably resulted from subglacial meltwater abrasion. The striations vary only slightly in orientation across the pit floor, and have an average trend of 167°. This trend is consistent with the east-northeast orientation of local end moraines. Crescentic marks are associated with the striations (Figure 3). Some of these fractures are convex in the direction of glacial flow, while others are concave. An unusual "channel" trends southwest across the bedrock surface near the pit entrance ramp. The origin of this feature is uncertain and may have been a composite of several processes. Some parts of the channel walls are meltwater-abraded, while others are striated or appear to have been plucked along bedrock joints.

Unit 2 - end moraine deposits. The two buried moraines exposed by pit operations in 1992 trend E-W to ENE-WSW across the bedrock surface. (Additional moraine deposits have since been exposed elsewhere in the pit complex). The southern moraine varies in composition from massive glacial diamict (till) with sand laminae, to a complexly sheared and interlayered mixture of till, sand, and gravel (Figure 4). The diamict lithofacies is olive-gray, sandy, stony, non-fissile, and moderately compact. It may include both flowtill and lodgement till. The till contains stones up to 3 m in diameter, some of which are faceted and striated. The layered parts of the moraine resulted from minor glacial readvance and mixing of till with proximal submarine fan deposits, in the manner described by Smith and Hunter (1989). This tectonic layering dips toward the proximal side of the moraine, and together with thrust faults and recumbent folds, indicates ice shove from the north. The northern moraine is less clearly defined, and consists chiefly of sand and gravel with a few diamict lenses. The latter sediments were deposited at the ice margin as proximal submarine fans, and are further described under Unit 3 below. The trough between the two moraines is conformably filled by laminated silt and clay of Unit 5.

Unit 3 - proximal submarine fan deposits. There is overlap in the depositional environments of Units 2 and 3, and the distinction between them is not always clear in the field. Proximal fan deposits form a major component of some moraines in the area, but they also can occur independently. These fan deposits were emplaced where high-energy meltwater streams emerged from the mouths of subglacial tunnels at the glacier margin. In the Toppi Pit, they are composed chiefly of massive to well stratified pebble-cobble gravel and sand. Bedding dips at shallow to moderate angles, and locally exhibits folding and reverse faults due to ice-shove. The coarsest fan gravel in the northern moraine (mentioned above) seems to occur as mound-shaped concentrations in the lower part of the section. These gravels may be tunnel-mouth deposits. The gravel fraction in this moraine is poorly sorted, massive to weakly stratified, mostly angular to subangular, and clast-supported. The proximal fan sediments higher in the moraine are stratified to a greater degree, with lenses of planar-bedded sand. Contacts between Units 3 and 4 range from gradational to sharp and unconformable.

Unit 4 - distal submarine fan deposits. This unit is generally less than 3 m thick, and is not present in all parts of the pit area. It marks a transition between Units 3 and 5 as ice retreat resulted in lower-energy sedimentation from density underflows and settling of mud from overflow plumes. Distal fan deposits in the Toppi Pit consist of well-stratified, interbedded sand, silt, and clay. They show sub-horizontal to gently-dipping planar beds, locally offset by normal faults and convoluted by water-escape structures. Channel-shaped unconformities occur at the base of this unit, within the unit itself, and even along the

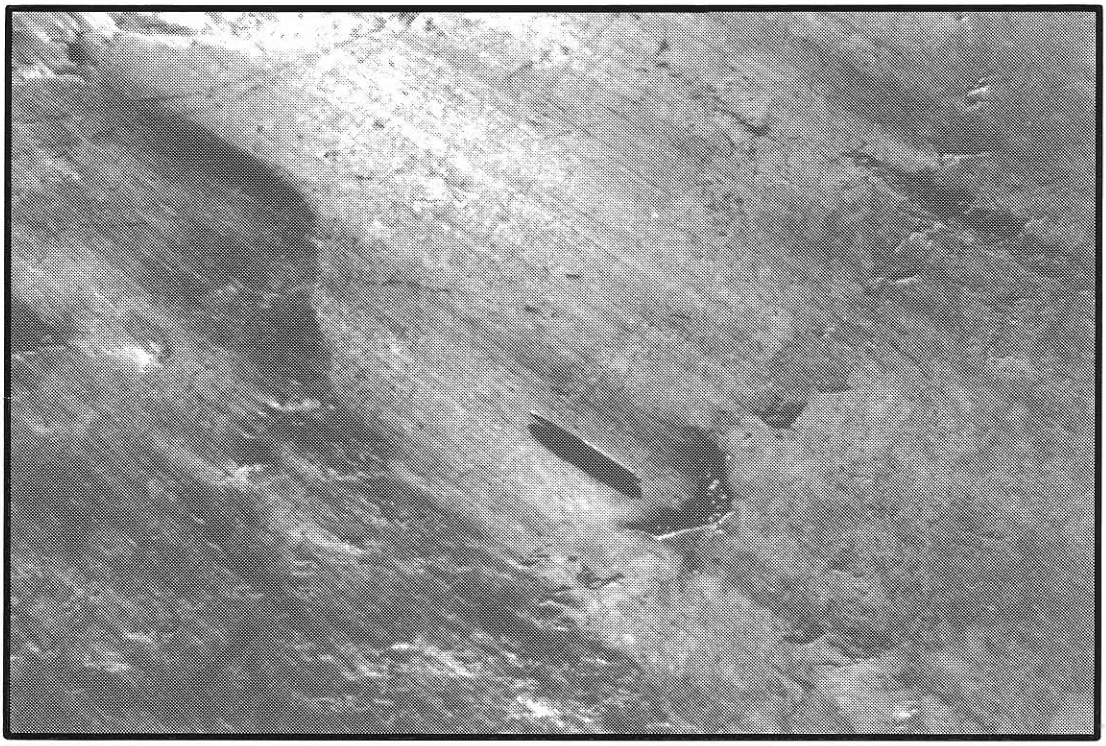


Figure 3. Crescentic mark on glacially polished and striated bedrock in floor of Toppi Pit. View to southwest.

Figure 4. Section through end moraine in the Toppi Pit, showing sheared mixture of diamict, sand, and gravel. Note shovel for scale; photo taken in 1992.

overlying contact with Unit 5. The channel-fills within the basal and internal channels locally include rip-up clasts of clay derived from erosion of the channel walls. Some of the sand beds in the channels are normally graded; others appear massive. These channel features are inferred to have resulted from submarine slumps and scouring by turbidity currents. Contacts between Units 4 and 5 range from gradational to sharp.

Unit 5 - glaciomarine mud. This unit consists of laminated silt, clay, and minor very fine sand (Figure 5). It was deposited in a quiet-water environment on the sea floor, and is part of the regionally extensive Presumpscot Formation. The observed thickness along the north wall of the Toppi Pit in 1992 was about 6 m. *Portlandia arctica* shells recently collected in this part of the pit have a radiocarbon age of 10,375 ± 80 yr BP (AA-10159). This age is anomalously young, since the *Portlandia* shells are typical of ice-marginal environments in coastal Maine and often yield ages older than 13,000 yr BP. The sample in question was found to be contaminated by reprecipitated carbonate.

1995 UPDATE

There are now three levels in the northern part of the pit, with most excavation activity in the expanded intermediate level. The "southern moraine" described above has been largely removed, leaving only scattered boulders on the bedrock floor. Fan deposits and overlying Presumpscot Formation are still well exposed in the high face on the north edge of the pit. Sections in the east-central part of the pit expose a third moraine, with the lower cut showing bouldery till with sand lenses, and the upper level showing diamict (flowtill?) lenses interbedded with sand and gravel. The morainal sediments are overlain by glaciomarine mud (Presumpscot Formation).

On the south side of the pit, recent digging has exposed sections in yet another moraine. From bottom to top, the general sequence consists of: (1) bouldery till, which is locally a very stony rubble; (2) proximal and distal submarine fan deposits; and (3) laminated glaciomarine silt, clay, and sand (Presumpscot Formation).

Questions:
1. What glacial processes formed the large-scale erosional features on the bedrock surface?
2. Can we distinguish ice-tunnel deposits in the present exposures?

5.40	0.10	Depart Toppi Pit. Turn L on Methodist Rd. and return to Westbrook.
8.70	3.30	Continue straight at light.
9.70	1.00	Turn R onto Business Rte. 25 (W).
9.90	0.20	Bear R on Rte. 25 and proceed W to Gorham. Go straight through center of town.
14.30	4.40	Just past jct. with U. S. Rte. 202 (S) in Gorham, turn L onto Flaggy Meadow Rd.
18.10	3.80	Go through jct. at Groveville, veering slightly to L onto Groveville Rd.
18.70	0.60	Turn R onto gravel rd. leading into Groveville Gravel Pit.

Figure 5. Glaciomarine mud filling depression between moraines in the Toppi Pit. Photo taken in 1992.

STOP 2: GROVEVILLE GRAVEL PIT - End moraine and glaciomarine fan (Standish Quadrangle)

Leader: John Gosse

This site provides a well-exposed complete vertical section through an end moraine and a submarine fan. As we approach the gravel pit from the northeast, pay attention to the three distinct types of glacial landforms around Groveville and Buxton Center (Figure 6). The gently rolling terrains with linear ridges are clusters of *end moraines,* which may be draped by a veneer of the sandy to clayey glaciomarine Presumpscot Formation. These moraine clusters locally are up to 1 km across. The elevated flat regions that appear well-drained are the tops of gravelly and sandy glaciomarine *fans* and *deltas*. Fans have well–bedded foresets but lack the overlying topset beds that characterize deltas. The flat sandy surfaces are typically not the original tops of the submarine deposits, but are the product of wave reworking during the recession of the postglacial sea. The poorly drained, flat to gently sloping regions are variably thick units of glaciomarine clay and silt of the Presumpscot Formation (Bloom, 1960).

From geologic evidence in and around the Groveville area, including a reasonably precise estimate of the local limit of the marine incursion which accompanied ice-marginal retreat, we can confidently rule out that these deposits are glaciolacustrine. The exposure in the Groveville gravel pit (Figure 7) is a superb site to show the relationship and association of the three distinct landforms mentioned above. The following description and interpretation is based on a fresh pit face (trending ~ 160°) visited in 1990 and 1992. From north to south, the section reveals a coarse glacial diamict (comprising the end moraine), which interfingers with a cobble gravel to sand unit (interpreted to be a subaqueous fan), which in turn interfingers with (and in places appears to grade into) a sandy (recessional) facies of the Presumpscot Formation. There appears to be a definite fining southeastward (toward ~ 100°) within the latter two units.

The formation of end moraines in Maine has been discussed by Smith (1982) and Smith and Hunter (1989). The general model is that the moraines formed in front of a grounded ice margin or under a floating margin at the grounding line. The end moraine at Stop 2 is a narrow, low-relief ridge which extended laterally for >60 m before mining. The end-moraine diamicton here consists of a non-compact, massive, clast-supported, poorly sorted cobble gravel with rounded clasts up to 40 cm in diameter. Striated clasts are uncommon but do exist. The coarse nature of the deposit makes identification and interpretation of any sedimentary structures difficult. It is uncertain if all of the diamict at this stop was deposited *in situ* at the ice margin, but there is no evidence to suggest significant reworking from wave action during recession. At another section through an end moraine (~ 2 km south of here) the diamict is matrix-supported, and a pebble fabric analysis (Figure 8) supports the depositional model in which debris sliding is the dominant process, rather than meltout or till lodgement.

The presence of subaqueous fans on the seaward flanks of end moraines is very common below the marine limit in Maine. At Stop 2 the fan grades from a cobble gravel to fine sand over a distance of only 20 m. The fan is characterized by thin, laterally extensive sequences which fine upward, outsize clasts that may be dropstones, and large and small fluid injection structures which may indicate the rapidity of deposition. This fan forms a wedge along the distal edge of the moraine, although it appears that some fans in the Standish region are conical and radiate from point sources on the moraines. In places at Stop 2 it

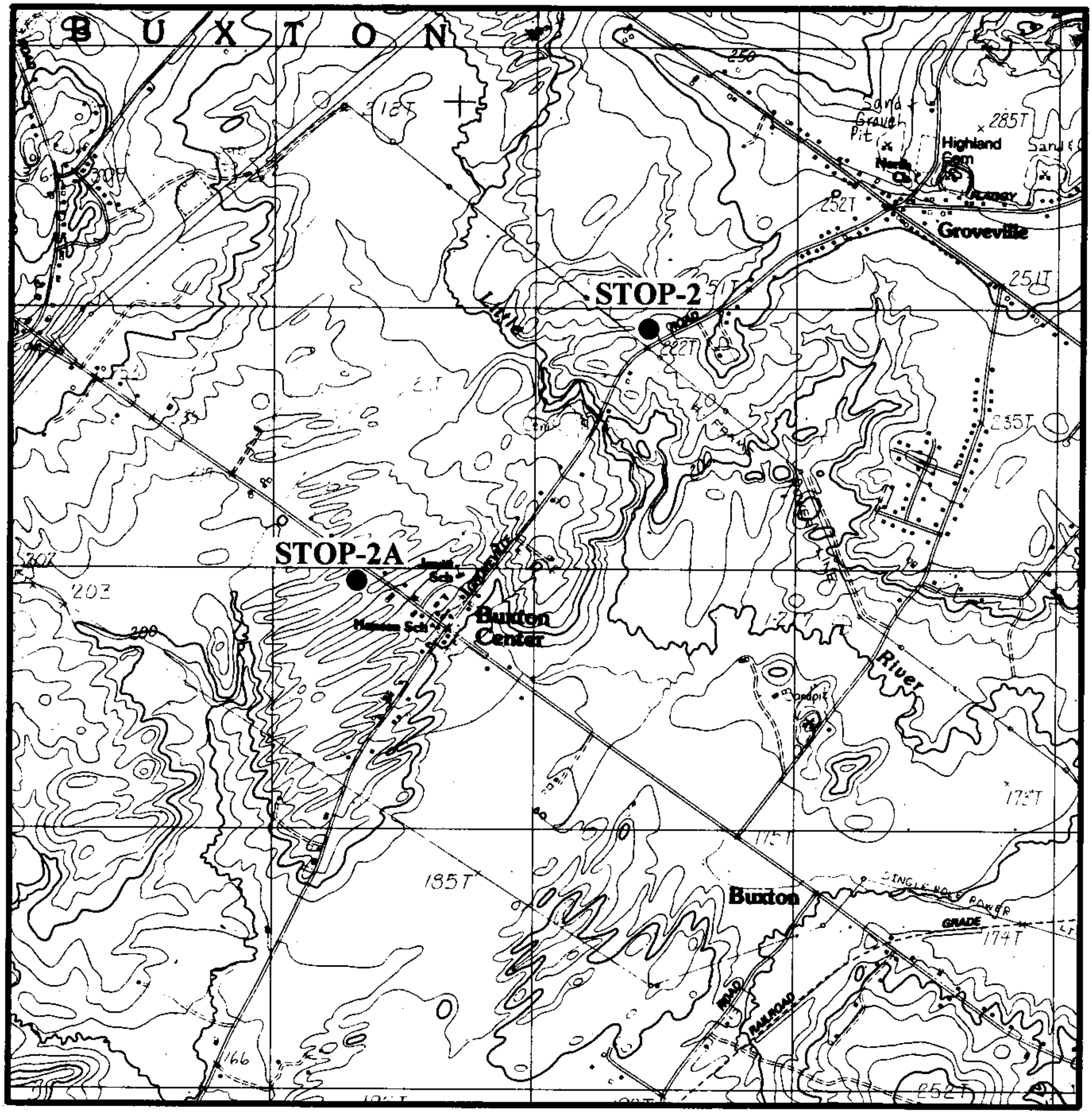
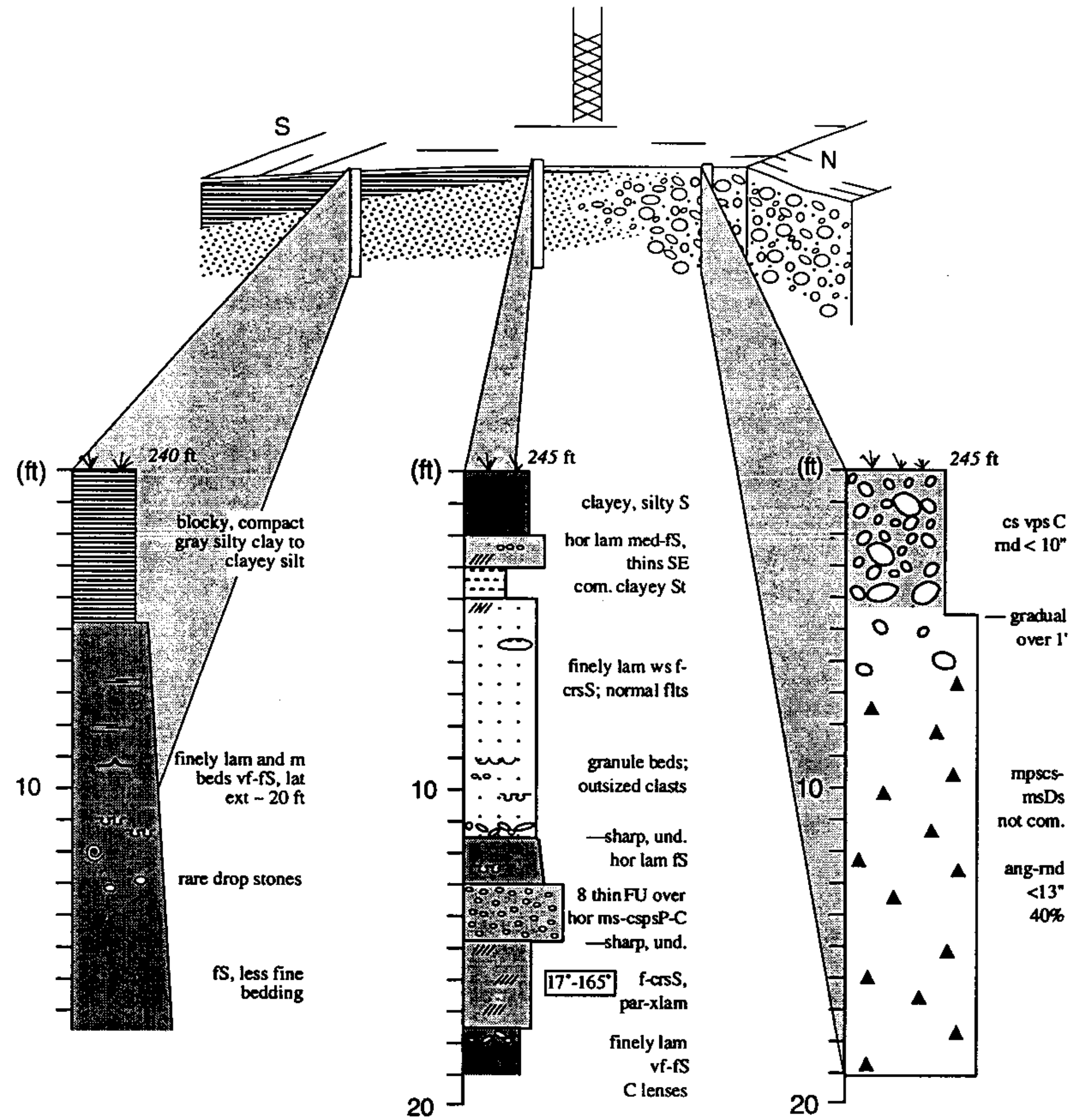


Figure 6. Location map for Stop 2 (Groveville Pit) and Stop 2-A (Buxton moraines).



asymmetric ripples

symmetric ripples

/// parallel cross laminations

load casts

convoluted bedding

255 ft Elevation of section top (feet asl, taken from topo map)

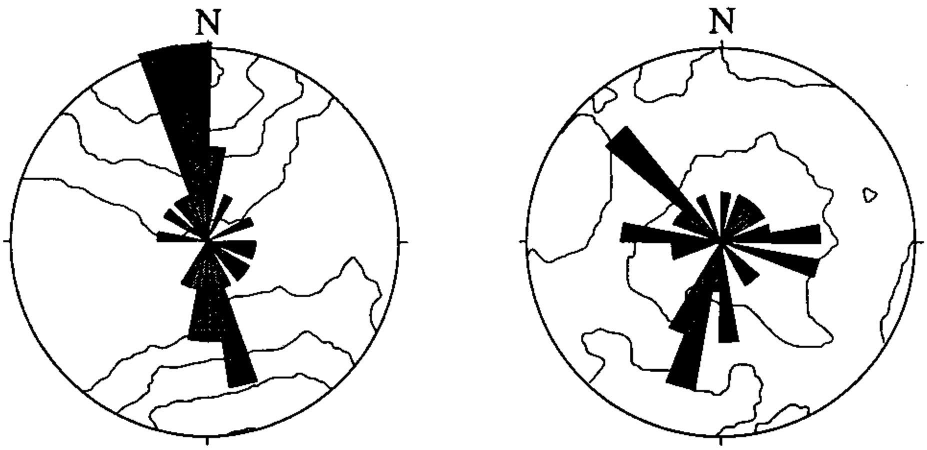
$\boxed{20°\text{-}206°}$ paleoflow (dip-trend)

original soil profile

fluid escape structures

rnd-rounded	ext-extent	
sph-spherical	ms-matrix supported	par-parallel
hor-horizontal	cs-clast supported	esc-escape
lam-laminations	m-massive	pebb-pebbly
xbed-cross bedded	ps-poorly sorted	cobb-cobbly
lat-laterally	com-compact	

Figure 7. Schematic sections of stratigraphic units exposed in the Groveville Pit.



n = 30
$S_1 = 0.64$

n = 30
$S_1 = 0.47$

ST-90-083
deformed lodgement till

ST-90-013
end moraine

Figure 8. Combined rose and contoured (Schmidt equal area, Kamb plot) diagrams of pebble fabric data from two separate till localities in the Standish region. The lodgement till has a strong fabric that parallels the local ice flow direction. The end moraine has a more random clast orientation.

appears that the fan and moraine units are interbedded, indicating the two formed contemporaneously. Admittedly, however, it is very difficult to distinguish one from the other. Because of these spatial and temporal associations and their size, end moraines and fans have commonly been mapped as one unit.

Farther seaward, the fan unit appears to be interbedded with and overlain by compact, blocky, gray fine sand, silt, and clayey-silt beds of the Presumpscot Formation. We do not interpret these fine-grained deposits to simply be distal foresets or bottomsets of the fan for two reasons. First, the Presumpscot Formation sediments appear to have a considerable clay content, suggesting they were deposited by flocculation. Second, the package fines upward, implying that the ice margin had continued to retreat, and does not coarsen upwards as would be expected in a prograding fan sequence.

Questions:

1. Could these units have formed somewhere other than at the ice margin?
2. Is there some way to determine how quickly this deposit formed?

Leave pit and continue SW on Groveville Rd.

19.60	0.90	Turn R (W) onto Rte. 22 at Buxton Center.
19.80	0.20	Park on side of Rte. 22 and note end moraines in field to S of road.

STOP 2-A: END MORAINES AND END MORAINE COMPLEXES (Standish Quadrangle)

Leader: John Gosse

In the southern half of the Standish Quadrangle several different glaciomarine facies cover the underlying lodgement till (Figure 9). Many of the end moraines in the Standish region are curvilinear till ridges that are laterally continuous for over 500 m. Where the end moraines occur as sets of closely- and generally regularly-spaced ridges (average crest-to-crest spacing is ~ 70 m), they are mapped as *end moraine complexes*. The moraines may be DeGeer moraines if they mark annual positions of a retreating ice margin (Smith, 1982; Solheim and others, 1990). The low relief of these ridges throughout the Standish area (mean height = 2 m; maximum studied = 3.5 m) is attributed to two factors: the truncation of the ridges by wave action during the offlap of the sea, and the ubiquitous thin veneer of Presumpscot Formation sand and clay (in places several meters thick) which broadens the crests and fills in the troughs. The marine sand and clay tends to bury the underlying till.

Before proceeding to Stop 3, we will detour to take a picturesque drive over two striking end moraine complexes around Buxton Center. End moraine complexes are prominent geomorphic features on topographic maps of southern Maine (Figure 6) and are even more spectacular on air photos. The majority of the moraines that you drive over are a massive and compact, clayey to sandy, matrix-supported diamict (locally called hardpan), but their composition is locally quite variable, even along the same ridge. Cemeteries are occasionally located where the crests of stratified moraines protrude through the marine clay veneer.

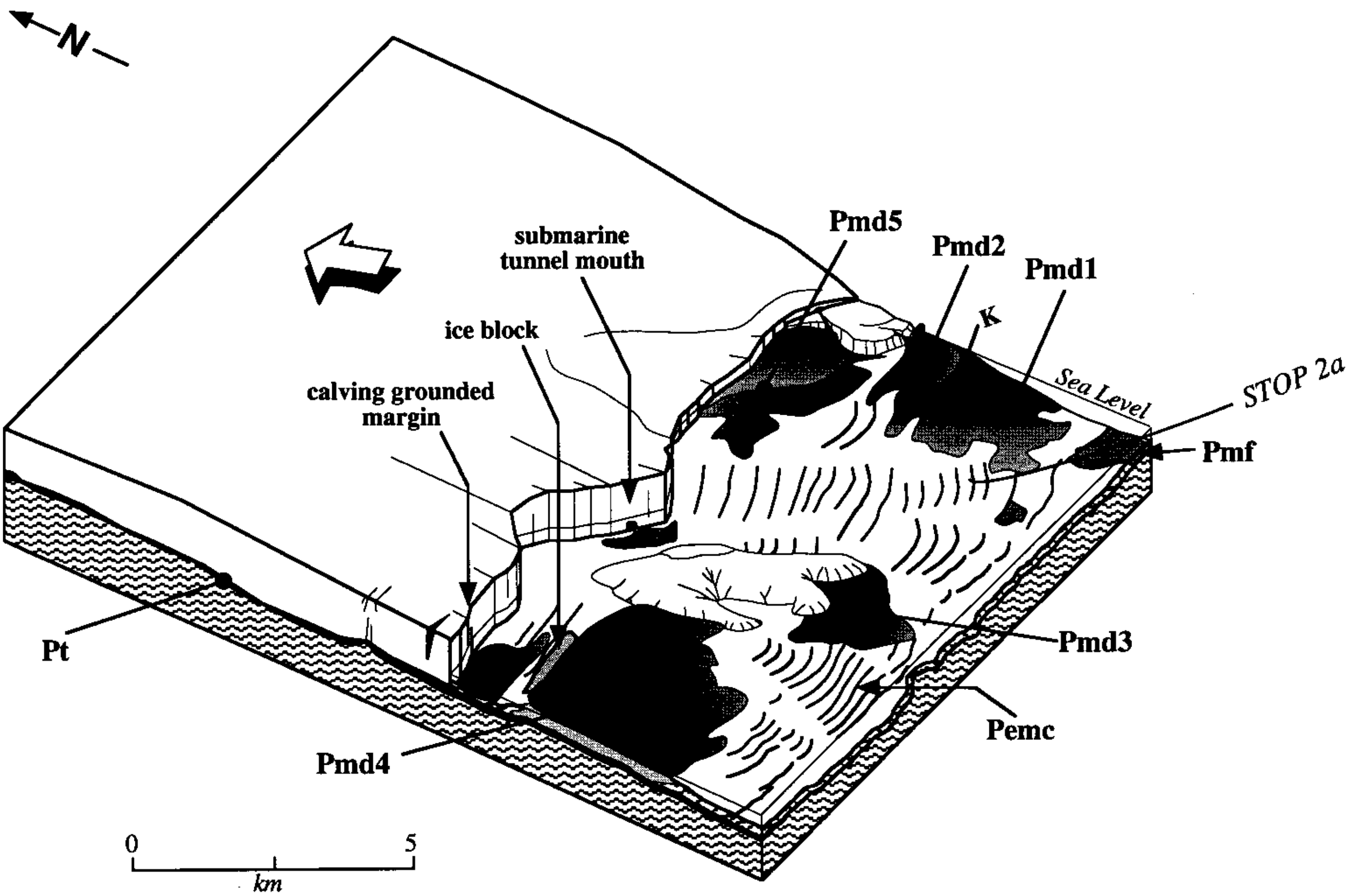


Figure 9. Schematic block diagram illustrating an ice-margin position and glacial sedimentation during the retreat of the Wisconsinan ice sheet in the Buxton area. Shadowed arrow indicates retreat transgressive relationships, i.e. before major reworking during sea regression. Ice margin is mostly grounded but is calving in places. Units **Pemc** and **Pmf** are an end-moraine complex and submarine fan, respectively. **Pmd** numbers indicate specific glaciomarine deltas mapped in the Standish Quadrangle. **T** = delta plain (topset unit); **K** = kettle. Presumpscot Formation not shown, but would cover most of the former sea bottom.

Some of the end moraines and associated ice-margin indicators (e.g. multiple kettle holes and ice-contact deltas) can be loosely connected across the entire area to indicate the position and geometry of the contemporaneous ice margins. Their connectivity implies that ice retreat through the Standish area was systematic and at least some of the ice remained active during its recession (Leavitt and Perkins, 1935; Borns and Hagar, 1965; Borns, 1967; Koteff and Pessl, 1981).

Keep in mind that where postglacial trellis and dendritic drainage has incised marine and glaciomarine sediments, the resulting trough-crest morphology closely mimics the morphology of the end moraine complexes, and this topography may be easily mistaken for moraines.

Question:
Are these end moraine complexes DeGeer moraines?

Continue W on Rte. 22.

22.80	3.00	Cross Rte. 112.
24.00	1.20	Turn R onto Rte. 35 (N).
27.80	3.80	Jct. in Standish village. Continue straight (N) on Rte. 35.
30.00	2.20	Sebago Lake village. Continue on Rte. 35.
36.90	6.90	Jct. in North Windham. Turn R (S) onto U. S. Rte. 302.
38.70	1.80	Turn L (at sign for "Moore's Motor Carriage") and follow gravel road to where it ends in the Tandberg Pit.

STOP 3: TANDBERG PIT - Glaciomarine delta and fossil plant locality (North Windham Quadrangle)

Leader: Woodrow Thompson

The Tandberg Pit (Figure 10) is located in the southern part of a glaciomarine delta complex (Bolduc and others, 1994). There are several coalescent deltas in this complex, covering much of the area around the town of North Windham. As we drove south from town on Route 302, we traversed the distal part of one delta plain, then quickly rose to the 320-foot contour at the proximal margin of an older delta. A third delta was crossed still farther to the south, just before turning onto the pit access road. Meltwater drainage from at least two tunnel (or open channel) systems delivered sediment to the North Windham delta complex. One system is marked by eskers north of town, on the east side of Route 302. The other is indicated by an esker at the south end of Little Sebago Lake. Esker feeders for the earlier-formed deltas probably are buried beneath the delta complex.

The contacts between topset and foreset beds in the North Windham deltas are not well exposed, so it is difficult to make a precise determination of late-glacial relative sea level in this area. Thompson and others (1989) measured an elevation of 300 ft (91.4 m) on a possible topset/foreset contact in the northeast part of the Tandberg Pit, but the "topset" gravels may have been reworked by wave action. Marine erosion of delta tops in southwestern Maine commonly has produced shoreline and nearshore deposits that can resemble topset beds, especially in a poor-quality exposure. The contact between these deposits and the underlying

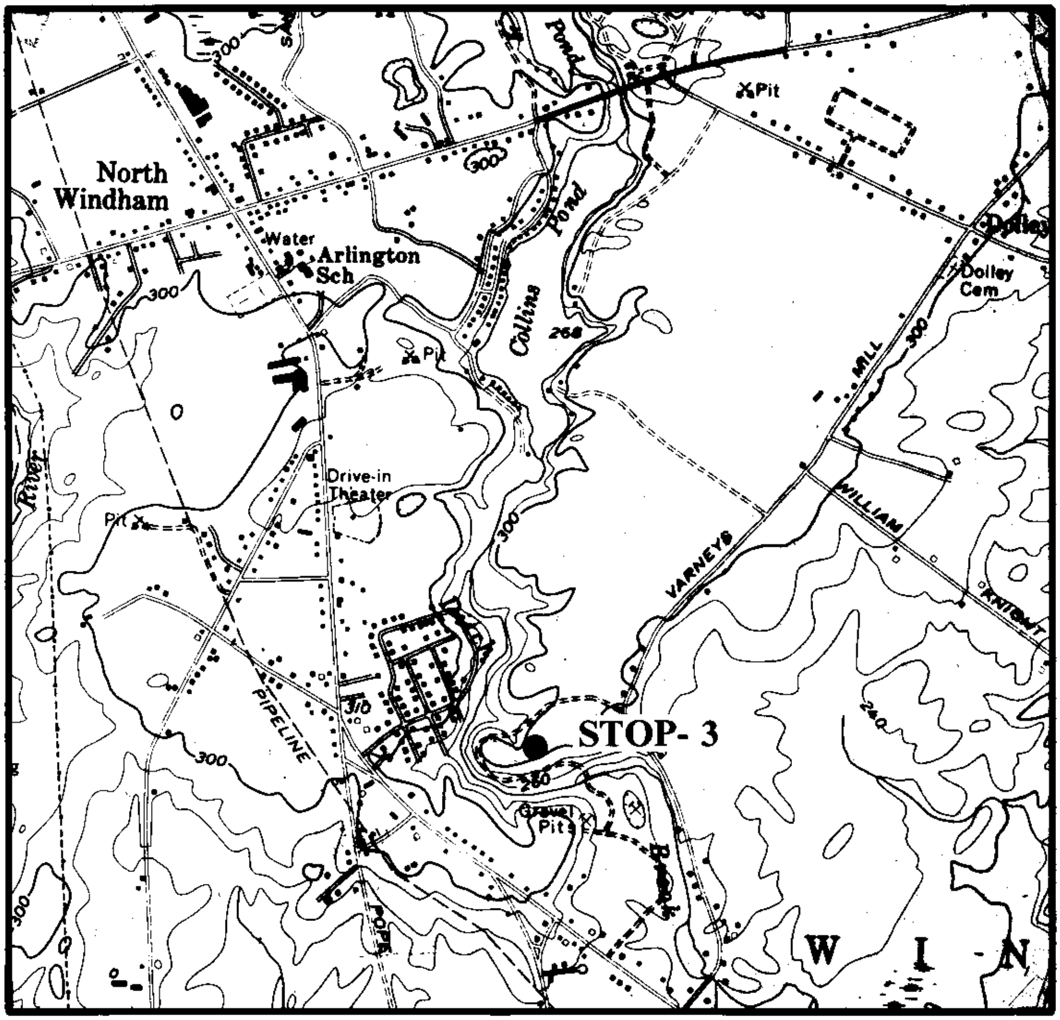


Figure 10. Location map for Stop 3 (Tandberg Pit, North Windham).

eroded foresets is often somewhat lower than the original topset/foreset contact. A section in the Canal Road delta, 2.5 km WSW of this stop, indicates an upper marine limit of at least 307 ft (93.6 m); and the regional pattern of marine-limit contours suggests a paleo sea level of about 310 ft (94.5 m) at North Windham (Thompson and others, 1989).

Aerial photographs taken before this pit was opened show kettled ice-contact topography. Remnants of a kettle can be seen in the woods north of the pit, and collapse structures are present within the pit area. Only the highest (ENE) part of the pit intersects the top of the delta. Most sections have exposed either foreset beds or earlier ice-contact deposits. The latter may include moraines or other ice-marginal deposits that were buried by the prograding delta. Several exposures in the western part of the Tandberg Pit showed interbedded silt-clay (mud), sand, and peat overlying the glacial sand and gravel. These younger sediments contain abundant organic material, ranging from twigs and leaves in the lower part to peat in the upper part. The muddy facies resembles the marine Presumpscot Formation, which laps onto the delta front just southeast of here. However, radiocarbon ages of the plant remains imply a freshwater origin. No pond could have existed here in recent years because the delta margin has been breached by Ditch Brook, but this incision may have occurred historically due to human intervention (see below).

Two interesting exposures on the north side of the pit ("West Section" and "East Section") were documented for this guidebook in late 1994. They were approximately 50 m apart, and separated by a concealed interval, but the organic-rich sediments in both sections were deposited in the same kettle basin. However, the scheduled reclamation of the pit proceeded faster than expected, and these sections were graded over shortly before the Friends of the Pleistocene trip. The stratigraphic descriptions are retained here because of their significance to the late-glacial and postglacial history of the region. In place of these sections, we will examine a small opening just west of the former West Section, showing the fossiliferous mud overlying sand and gravel.

The following units were exposed in the **West Section**, which was a small east-west face. Unit 1 is oldest.

Unit 1 - Gravel. The lowest unit in the section is a sandy, poorly-sorted, pebble-boulder gravel. The base of this unit is not exposed. Stones in the gravel are mostly subangular to subrounded. The contact with Unit 2 is sharp, conformable, and dips eastward at 22°. This unit is interpreted to have formed in an ice-contact environment. It may be part of a submarine fan or ice-tunnel deposit that was buried by the overlying sediments.

Unit 2 - Gray silt. This unit consists of dark gray (5Y-4/1) clay-silt-sand. It is 30 cm thick, with thin planar beds that dip eastward at 25°. The sediment is silty in the lower part and fines upward to clay-rich mud with sand laminae. (Previous exposures at this site also showed an additional unit between Units 1 and 2, consisting of about 40 cm of well-bedded pebbly sand that probably washed into the kettle from adjacent glacial sediments.) A few dropstones were noted in Unit 2, and it contains fine-grained organic material of terrestrial origin. A sample of *Dryas integrifolia* leaves yielded an AMS radiocarbon age of 10,550 +/- 90 yr B.P. (OS-4415). This age falls within the Younger Dryas climate interval. Unit 2 was deposited in a ponded environment, and it is inferred from the radiocarbon age (which postdates the marine submergence) that it is a freshwater deposit.

Unit 3 - Brown silt. Unit 2 is conformably overlain by 22 cm of grayish-brown (2.5Y-5/2), well-bedded sandy silt. It is differentiated from Unit 2 mainly by its brownish color, which is attributed to an abrupt increase in the influx of fine organic debris. This unit also contains macroscopic plant remains, including flattened twigs. The twigs resemble those which were collected for dating in the East Section, and thus are inferred to be *Populus balsamifera* (poplar). Based on the age of Unit 2, Unit 3 is likewise nonmarine.

Unit 4 - Sand and plant debris. Unit 3 consists of 120 cm of stratified sand with a few pebbles. It conformably overlies Unit 3, and is distinguished by its coarser texture and abundance of plant macrofossils. The sand is thought to have been derived from erosion of delta foreset beds adjacent to the kettle.

Unit 5 - Peat. This unit consists of matted peat, with a much higher content of plant material than the underlying units. It is at least 40 cm thick, and may have been thicker, but disturbance from the pit operation has obscured the original ground surface.

The **East Section** (Figure 11) was a long NE-SW pit face that exposed four principal units:

Unit 1 - Sand and gravel. The lowest unit in the section consists of delta foreset beds. These beds range in texture from sand to boulder gravel (mostly sand in upper part), and dip up to 20° to the WNW. The bottom of the unit is not exposed. The upper contact, with Unit 2, is conformable and gradational over an interval of 20 cm. Structures in Unit 1 include contorted bedding, and both normal and reverse faults. This deformation resulted from slumping on the delta front and/or collapse adjacent to melting ice. Oversteepened gravel foresets high in the northeast end of the section suggest ice-contact collapse.

Unit 2 - Silt. This unit is 84 cm thick, as measured from the base of Unit 4 down to the gradational contact with Unit 1. It is composed of dark gray (5Y-4/1), well-bedded clay-silt with sand laminae. Dropstones are present, and fossil wood fragments occur in at least the upper part (down to just above the tapering intertongue of Unit 3). Flattened *Populus balsamifera* (poplar) twigs are the dominant macrofossils, and one of the stratigraphically lowest twigs (25 cm below top of unit) has been dated. The radiocarbon age of the sample is 12,100 +/- 110 yr B.P. (OS-4416). Relative sea level was falling in southwestern Maine by this time, so the plant remains are assumed to have been deposited in a nonmarine environment.

Unit 3 - Sand. A wedge-shaped sand lens intertongues with Unit 2, and pinches out within the latter unit in the down-dip direction (to the southwest) (Figure 12). This unit is up to 80 cm thick in the central part of the face. It has large-scale planar cross-bedding that dips westward at 26°, and there is some evidence of internal slumping. The sand is interpreted as a nearshore deposit derived from erosion of the adjacent delta foresets.

Unit 4 - Interbedded sand-peat-silt. This composite unit sharply and conformably overlies Unit 2. It consists of at least 1.5 m of interbedded sand, peat, and silt (original ground surface has been disturbed). These sediments are part of the kettle-fill sequence, and appear to be correlative with Units 4 and 5 in the West Section.

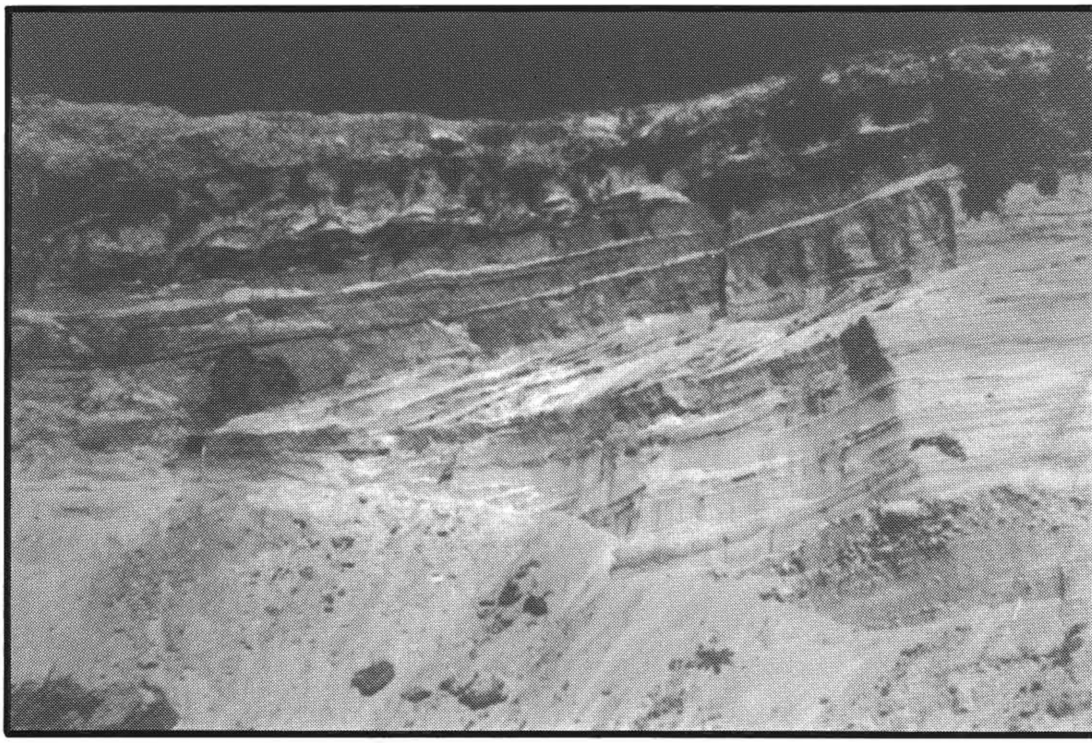


Figure 11. East section at Tandberg Pit. Pit face is 6 - 7 m high. View to northwest; photo taken in 1994.

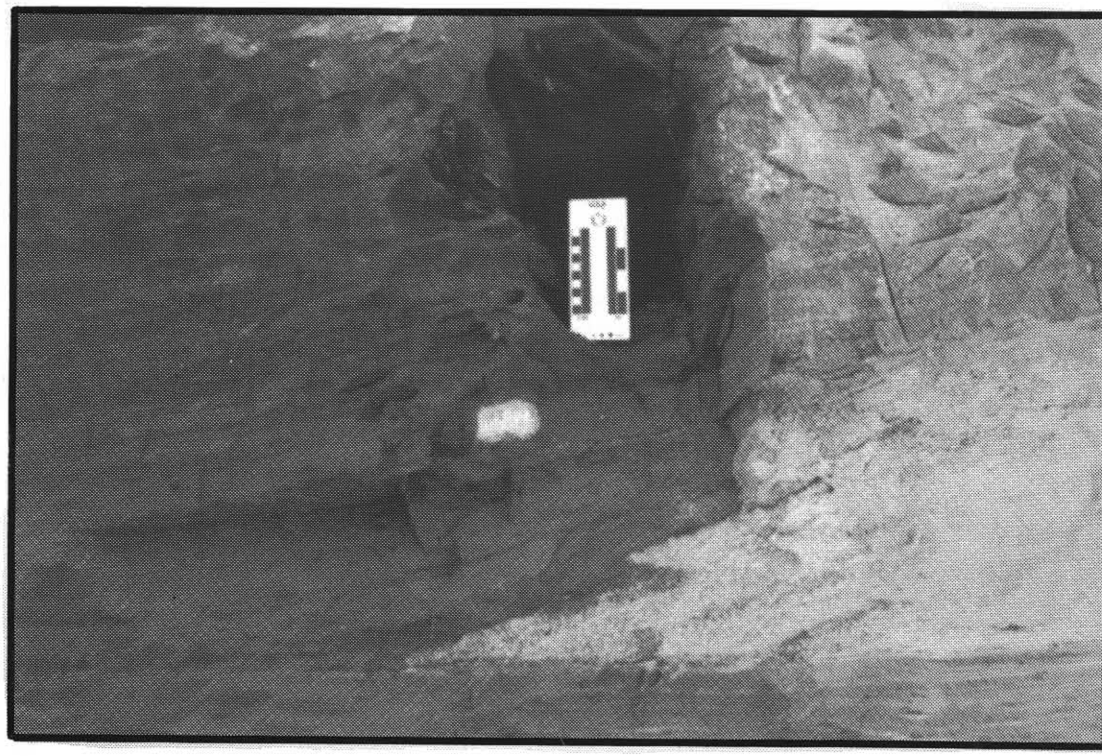


Figure 12. Close-up of left-central part of section in Figure 11, showing sand lens (Unit 3) intertonguing with mud (Unit 2). Scale card is graduated in centimeters and inches. Dated poplar twig (OS-4416) was located just above the card. Note dropstone in mud (center).

Historical records from the town of Windham reveal some fascinating details concerning Ditch Brook, which cuts through the delta complex and passes the west edge of the Tandberg Pit. This stream drains a nearby chain of ponds, which in turn receive the outflow of Little Sebago Lake to the north. Little Sebago formerly drained westward into Sebago Lake via Outlet Brook, but a new outlet was excavated at the head of Ditch Brook to divert the drainage southward. Smith (1873) described the disastrous consequences resulting from tampering with the drainage system and building a dam on a sand-and-gravel foundation:

> "Col. Edward Anderson caused an artificial outlet to be made from the south end of the third pond into Smith's Brook [= Ditch Brook], which empties into Pleasant River, for the purpose of supplying the mills on that river with an additional quantity of water. This outlet increased in size, until two sawmills were erected upon it. June 4, 1814, the water undermined the mill dam, swept it and the mills from their foundations, disrupted the bed of the stream ... and forced its way into Pleasant River, a tributary of the Presumpscot. In a few hours the outlet was increased 50 feet in depth and 200 feet in width. ... The sudden eruption of this great body of water carried away one saw mill, one grist mill, and four bridges on Pleasant River, and Gambo and Mallison Falls bridges on the Presumpscot, overflowed the intervale and low lands, caused the water to run up Pleasant River and the tributaries of the Presumpscot; and caused many who saw the laws of nature thus reversed, and unacquainted with the cause, to believe the world was coming to an end instantly."

Smith added that a second dam was built in the same area, with even more disastrous results when it collapsed in 1861, triggering a flood that destroyed numerous mills and bridges! Dole (1935) noted that "...Ditch Brook has ever since been the principal outlet of Little Sebago; whereas before Anderson wanted water for his mill, that pond was many feet higher than it is now, as shown by traces of the old shore lines...". Although largely forgotten in modern times, these and other drainage modifications dating to the early 1800's have greatly altered the original postglacial landscape in parts of southern Maine.

Questions:

1. Were the plant remains deposited in a small kettle pond, or in a larger water body that extended northward and perhaps included present-day Collins Pond?
2. What was the source of the fossiliferous mud?
3. What was the origin of the high terrace (approx. elev. = 260 ft) seen near the pit entrance, and does this terrace indicate a historic/prehistoric drainage level for a former lake?

Go back to Rte. 302 and return to North Windham.

40.40	1.70	Jct. in North Windham. Continue N on U. S. Rte. 302.
50.70	10.30	Turn L onto rd. to Sebago Lake State Park.
52.30	1.60	Turn L into State Park and drive S to lake.
53.90	1.60	Park in lot for State Park beach and walk to lake shore.

STOP 4: SEBAGO LAKE STATE PARK - SONGO BEACH - The inland marine limit problem (Naples Quadrangle)

Leader: Robert Johnston

Sebago Lake State Park is situated along the boundary between Maine's coastal lowlands and central highlands (Denny, 1982). The coastal lowlands lie to the southeast with elevations up to 328 ft (100 m) above sea level, while the central highlands to the northwest range in elevation from 328 ft (100 m) to over 4000 ft (1200 m). Due to its location, Sebago Lake was an important transition zone during the final stages of late Wisconsinan deglaciation. The State Park is located along the north shore of the lake and is dissected by the Crooked and Songo Rivers (Figure 13). The Crooked River joins the Songo River at the present-day site of Songo Lock. Songo Lock opened in 1830 to carry raw materials and commodities from the Oxford Hills region to and from Portland.

Sand deposited by the Songo River during deglaciation, and sand and gravel deposited as a kame terrace, make up most of the surficial material found within the State Park. Numerous kettles and oxbow lakes are also present (Figure 14). The road leading from Route 302 into the park travels from an upland till surface (where we departed Rte. 302) onto a sand surface which was deposited by the Songo River during deglaciation. Note the flat to gently undulating land surface as we drive down the park road towards the beach. The Songo Beach picnic area is located on braided stream deposits. During the final stages of deglaciation the Songo River was a braided stream, and thick sheets of sand were deposited across the entire flood plain of the river (Bloom, 1960). A Maine Geological Survey seismic-refraction line, along the Thompson Point Road (Figure 13), shows a depth to bedrock of 133 ft (40.5 m). A nearby monitoring well has 23 feet of sand, over 28 feet of sand and clay, over 12 feet of sand, clay and gravel. Excellent exposures of the braided stream deposits are found along the banks of the modern Songo River, but we will not visit them on this trip.

Erosion of sand at Songo Beach in recent years has concerned park users and park personnel. In 1990, the Maine Geological Survey, in cooperation with the Maine Bureau of Parks and Recreation, initiated a beach profiling study to determine the causes of erosion on Songo Beach. Over four years of beach profiles have been analyzed to determine the dynamics of sand movement on the beach. To date no net loss of sand to the beach has been documented, but there has been movement of sand due to seasonal fluctuations of water level. Profiles at additional sites around the lake are presently being studied for insights into the causes of beach erosion.

The inland marine limit in Maine marks a sea-level highstand at approximately 14-13 ka. It was inferred to cross Sebago Lake near Frye Island by Thompson and Borns (1985) on the Surficial Geologic Map of Maine, dividing the lake into a northern glacial-lacustrine basin and a southern glacial-marine basin. In a study to examine the accuracy of the mapped marine limit in the lake we analyzed the nature of glacial-lacustrine and glacial-marine sedimentation in the lake basin. Recognition of the marine limit is usually based on mapped shorelines, glacial-marine deltas, and the distribution of glacial-marine sediments. We collected 20 sq. mi. (52 sq. km.) of side-scan sonar images, 31 mi (50 km) of seismic reflection profiles, and one piston core. Side-scan sonar records show coarse sand and gravel and extensive boulder fields at an inferred grounding-line position near Frye Island, where the marine limit was drawn. ORE

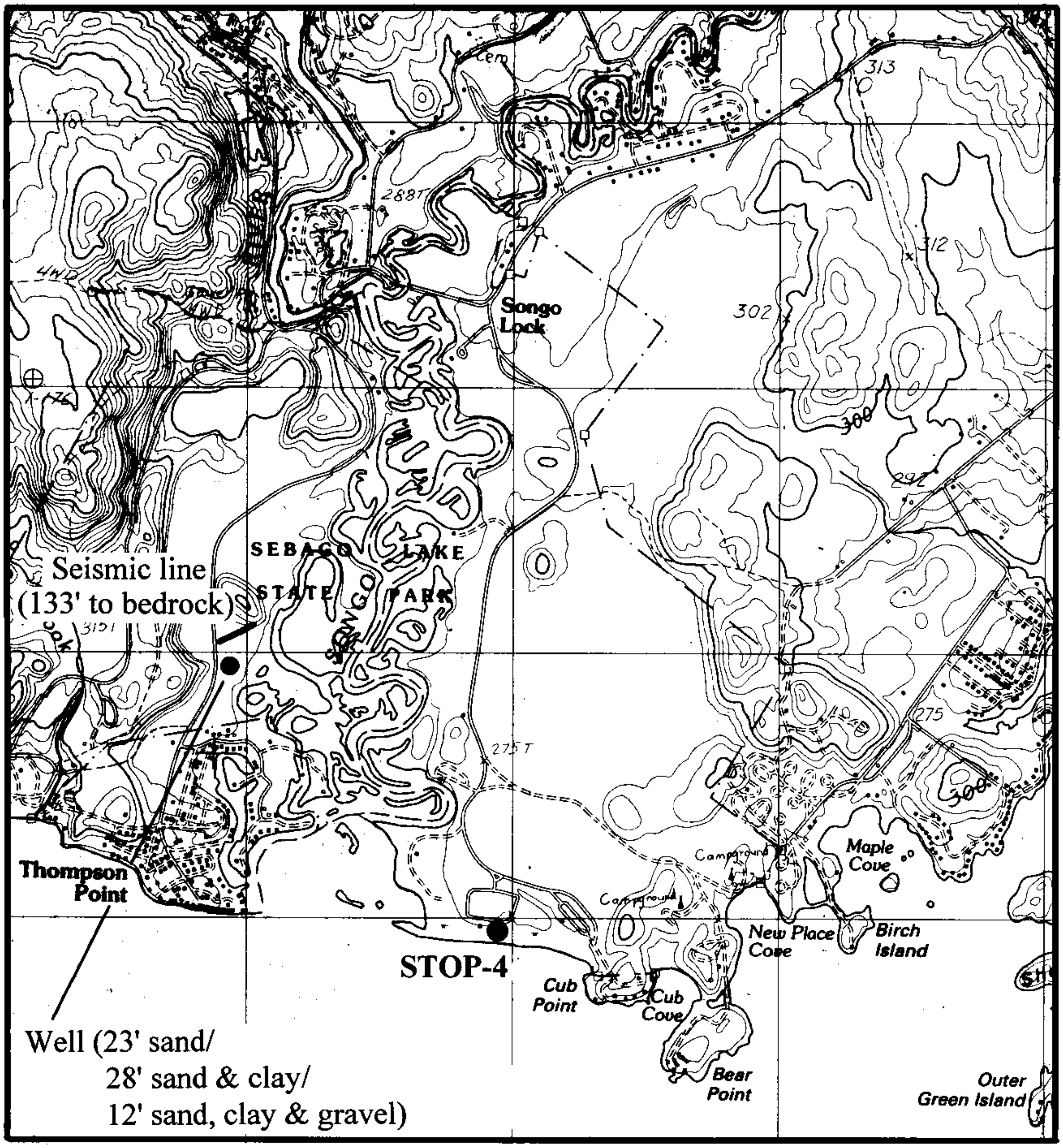


Figure 13. Map showing location of Stop 6 (Songo Beach, Naples) and nearby monitoring well and seismic line.

Figure 14. Aerial photograph, dated 10/26/63, showing the Songo River entering Sebago Lake. The extent of the Songo River delta, meander scars, and oxbow lakes are clearly seen.

Geopulse seismic reflection profiles reveal a basal draping unit similar to glacial-marine units identified off the Maine coast. Later channels cut more than 60 ft (20 m) into the basal stratified unit. In addition, till and a possible glacial-tectonic grounding-line feature were identified.

Paleoshorelines rim the nearshore basins, indicating a lower lake level 36 ft (11 m) below the present lake level. Slumps and possible spring disruptions are found in several locations. Slumping has carried fines into the deep basins, while wave-washed sand and gravel covers nearshore basins. The top seismic unit in the lake is an onlapping Holocene lacustrine unit. Total sediment thickness is much greater in the southern basin. The northern basin, > 97 m deep north of the marine limit, appears to have been occupied by an ice block. Retrieved sediments include 36 ft (12 m) of rhythmites, and sieving and total organic analysis of the core showed no microfossil/macrofossil evidence and little organic material.

Questions:
1. What might be the cause of the debris flows seen in the geophysical records of the lake floor?
2. What would have kept marine waters from entering the Big Basin?
3. What was the origin and base-level control of the kame terrace on the west side of the Songo River valley?
4. How could the postglacial lake level formerly have been lower than it is now, since the lake drains across a bedrock threshold?

55.60	1.70	Leave park, turning R on access rd., and return to Rte. 302.
57.05	1.45	Turn L onto U. S. Rte. 302 and continue NW.
59.40	2.35	Turn R onto Rte. 11.
60.50	1.10	Turn L and drive into P & K Sand and Gravel pit.

STOP 5: P & K SAND AND GRAVEL PIT - Glaciolacustrine sediments (Naples Quadrangle)

Leader: Robert Johnston

Stop 5, the P & K sand and gravel pit, is located approximately 7.5 km north of Songo Beach, on the flood plain of the Crooked River (Figure 15). The deposit of glaciolacustrine sediment exposed in this pit is the focus of our visit.

In its more active days, the pit exposed over 50 ft (15 m) of section (W. B. Thompson, personal communication). The oldest unit is an esker with a sand/boulder-gravel core, mantled by a thick section of well-stratified glaciolacustrine silt to very fine sand. The ground surface above the western side of the pit is flat and graded, at an elevation of ~ 320 ft (97 m). Along the east side of the pit, very coarse sand with fluvial cross-bedding was found overlying well-bedded silt, which in turn overlies the esker gravel. The sand unit in this pit is probably a stream-terrace deposit formed by the Crooked River.

The P & K Pit is presently being reclaimed, but in the past this deposit exhibited large-scale slumping in

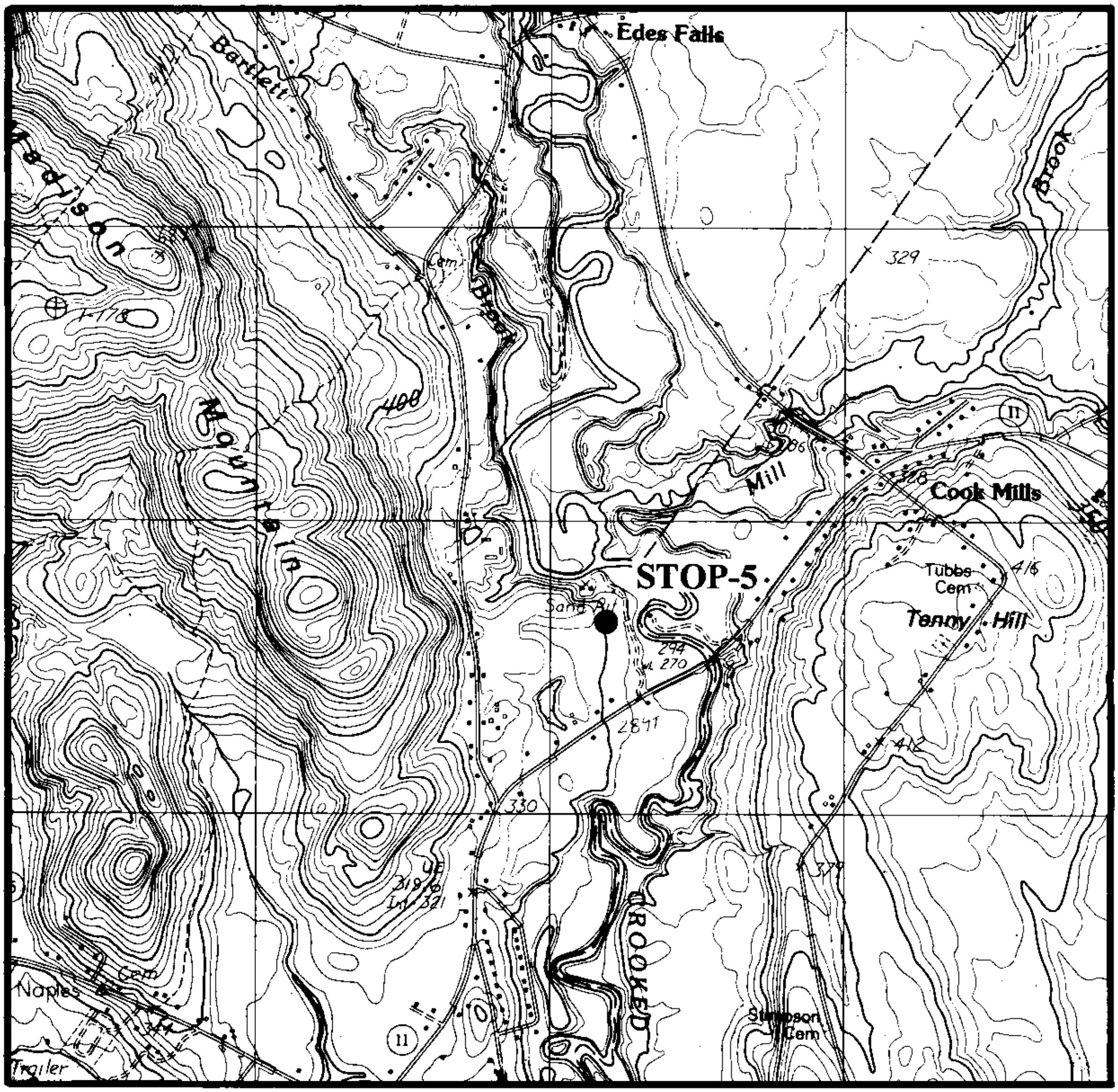


Figure 15. Location map for Stop 5 (P & K Pit, Naples).

the fan and stream terrace deposits (Figure 16), and thick accumulations of fine-grained glaciolacustrine sediment (Figure 17). The well-stratified lacustrine unit is characterized by repeated layers 3-5 cm thick, with additional laminations found within each layer. Sieving and macrofossil analysis of the glaciolacustrine sediment in the Pollen Lab at the University of Maine (C. Dorion, personal communication) yielded freshwater plant fragments and beetle parts. The elevation of the glaciolacustrine sediments (270 to 300 ft; 82 to 91 m) indicates that a drift dam or ice block must have been present downstream. One scenario would have been for a remnant ice block to have been present in the Sebago Lake basin, damming a late-glacial lake to the north. Subsequent melting of the ice would have allowed for the gradual downcutting of the Crooked and Songo River valleys to their present level. A second possibility would be to have a drift dam across a narrow constriction between the P & K Pit and the lake. Large-scale slumping of the lake beds may have been caused by sediment loading and/or seismicity related to the unloading of the crust by ice.

Questions:

1. Are the clay layers in the P & K Pit varves?
2. What is the origin of the sand units in the P & K Pit?

Return to Rte. 302.

61.65	1.15	Turn R onto U. S. Rte. 302.
63.00	1.35	Turn L (S) onto Rte. 11 in Naples village.
72.40	9.40	Turn R at East Sebago, staying on Rte. 11.
73.75	1.35	Park on R (W) side of Rte. 11 and hike through woods to bouldery ridges near road.

STOP 6: STONY BROOK ESKERS

Leaders: John Gosse and Woodrow Thompson

As you travel southward from Sebago Lake on Route 11, you will drive over sandy till and a lacustrine(?) delta complex at East Sebago. Just southwest of East Sebago village, there is a 0.5-mile long esker segment that trends parallel to the road on the east side. Notice that the adjacent hillslope is littered with large granitic boulders over 1 m in diameter. These boulder fields are typical of areas underlain by granitic plutons. The boulders were probably plucked from the granitic hills that were overriden, but because the boulder patches occur on both the stoss and lee sides of hills there has probably been some transportation. The boulder patches in this area may be a lag because they appear to be concentrated on hillslopes, and not on the horizontal till plains, suggesting that surface wash may have winnowed the finer sediment from the till surface. The upper marine limit in this area reached about 330 ft (100 m) above modern sea level, but it is unclear exactly how high it was (see Stop 7 for further discussion).

Stop 6 is situated near an esker ridge complex that extends to the head of Stony Brook on the east side of Route 11 (Figure 18). The ridges are comprised of well-sorted, massive to well-bedded sand and sandy

Figure 16. 1993 photograph of the P & K Pit, showing a sand unit on the western side of the pit with evidence of large-scale slumping.

Figure 17. 1993 photograph of rhythmically bedded glaciolacustrine sediments exposed in the P & K Pit.

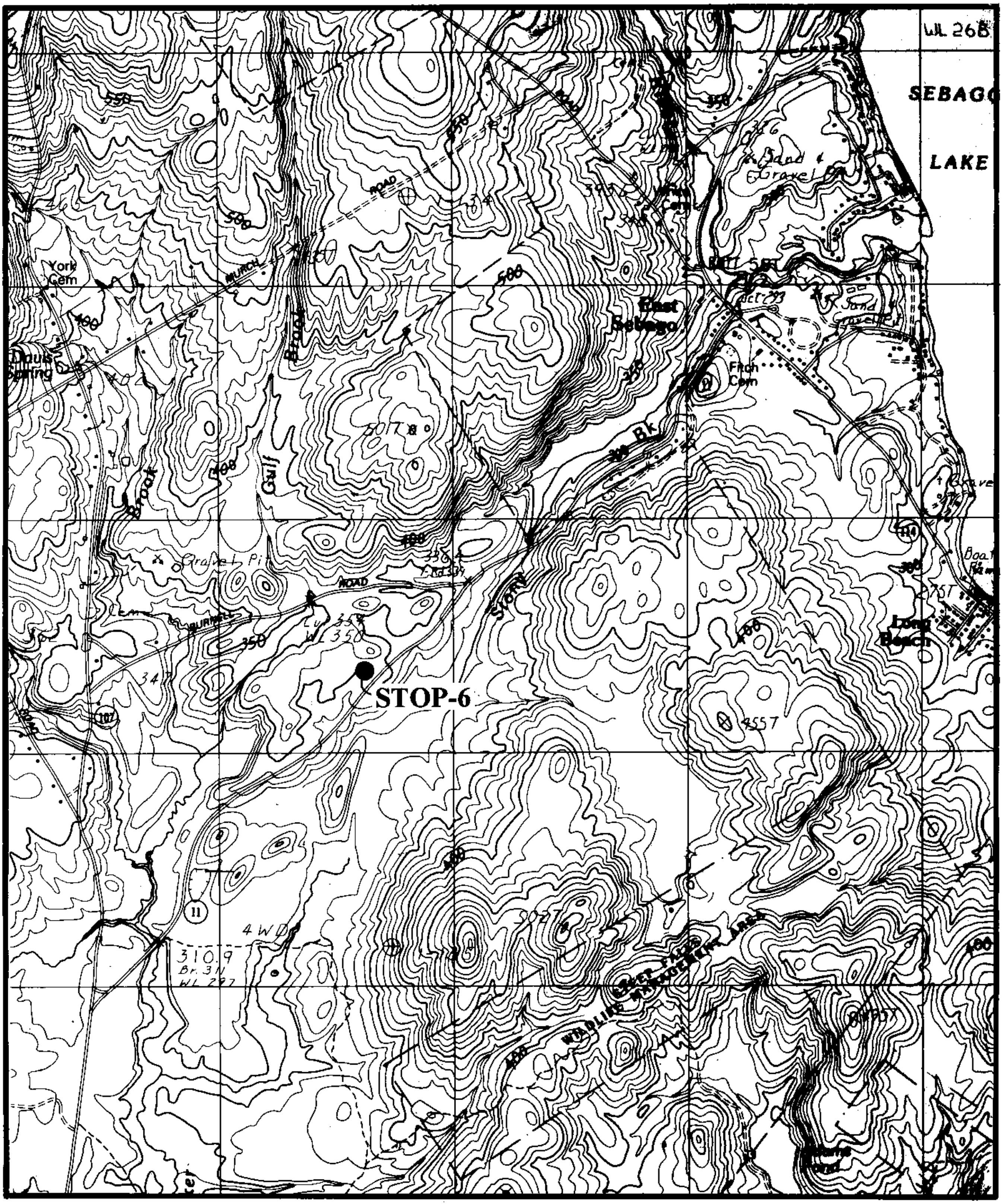


Figure 18. Location map for Stop 6 (Stony Brook eskers, Baldwin).

gravel with boulders up to 1 m in diameter, as revealed in two vertical road cuts. These eskers are well defined topographic features with local relief of 6-12 m, steep flanks, and narrow crests. Unlike other eskers in the area (e.g. Bonnie Eagle Esker, Standish Quadrangle) there is no evidence that wave action has altered the crests of these eskers.

Stop 6 investigates the bouldery landforms on the west side of the road that are interpreted as esker segments. These landforms are a less well-defined series of intermittent ridges with lower relief than the eskers on the east side of the road. The crest of at least one of the ridges near Stop 6 is sharp and narrow but generally broadens southward. There are conspicuous clusters of large boulders as great as 1.5 m in diameter on the ridges and on flat surfaces that surround the ridges. Based on the proximity to the neighboring unequivocal eskers, their predominantly sandy composition, their lateral extent, and shovel and auger probing into sorted sands and gravels, we believe this series of ridges is part of the esker system to the east. The flow direction of the glacial streams that deposited them is uncertain, but is inferred to be southwestward.

The location and southwest trend of these eskers may have implications for constraining ice flow direction. In the Steep Falls region, striae and other ice flow indicators record a southward (mostly 174° to 180°) and southeast (142° to 162°) flow. At two localities just south of Sebago Lake, ice flow appears to have been ~ 220°, implying that an ice mass may have flowed southwest from the lake basin, in agreement with other indicators of a Sebago Lake ice mass (previous Stops; Goldthwait, 1951). The eskers at this stop are parallel to this flow and may provide some constraint on the ice margin position and volume of this Sebago ice mass.

These ridges were formerly interpreted to be an erosional escarpment or a group of moraines — the controversy being due to the sandy nature of till matrix in the vicinity, the orientation of the ridges with respect to dominant ice flow directions (see below), the presence of the large boulders, and the lack of good vertical sections exposing their composition and structure. It is possible that the ridges are nested moraines comprised of loose sandy till (derived from the metasediments beneath and up-ice from the ridges). The presence of large boulders on the so-called 'till' is consistent with boulder patches on till in the surrounding area. The material beneath the boulder patches has not been examined. The orientation of the ridges is perpendicular to the southeast ice flow direction, concurring with the orientation of unequivocal moraines farther southeast.

Questions:

1. Are these ridges definitely eskers?
2. Was the orientation of these eskers controlled by the local topography or by ice flow from the Sebago Lake basin?

Return to road and continue S on Rte. 11.

75.00	1.25	Jct. with Rte. 107. Continue S on Rte. 11.
76.80	1.80	Turn R (W) onto Rte. 113 at East Baldwin.
77.75	0.95	Bear R onto Brown Rd.

78.20 0.45 Turn L into the Guptill Pit.

STOP 7: GUPTILL PIT - Baldwin Delta (Steep Falls Quadrangle)

Leaders: John Gosse and Woodrow Thompson

Simply put, we are still uncertain whether the delta at Stop 7 is glaciomarine or glaciolacustrine. An answer to this question would directly constrain the elevation of the upper marine limit in the Saco River valley, and has a major bearing on the nature of postglacial events farther up the valley (see Day 2 stops).

The Baldwin Delta has well-developed topset and foresets beds (Figures 19 and 20). The topsets are up to 2 m thick in places where a moderately well-developed soil indicates the delta surface has not been significantly excavated. The topsets range from well-bedded pebbly sand to clast-supported cobble gravel in which bedding is at best moderately developed. Rare cross bedding and weak imbrication within the topsets exposed in the east end of the pit indicate paleoflow toward the east. The gravels in the topset package seem to fine toward the southeast, from coarse moderately sorted cobble gravel to a better sorted pebbly sand.

The contact between the topsets and foresets is sharp and generally horizontal or gently dips 2° toward the northeast, but is undulatory in places where ice foundering is evident. The foreset package is at least 10 m thick. The foresets range from thick units (0.6 m) that fine upwards and consist of finely laminated pebbly sand to fine sand, to coarse structureless cobble gravel. Cross laminations, climbing ripples, and herringbone ripples have been identified within many foreset beds. The dominant dip of foreset beds in the pit sections is 26° toward 240° (WSW), but ranges from SSW to WSW, in an opposite sense to the paleoflow direction indicated by the topsets.

The clasts in the Guptill Pit average 20 cm or less in diameter (although a few boulders ~ 40 cm can be seen around the pit) and are almost entirely granitic (80%) or volcanic (20%). Many of the stones in the topset unit are subangular. The source of the sediment was probably the Sebago Batholith and volcanic rocks to the north, and not the surrounding Saddleback Mountains (which are formed of easily comminuted Silurian Rindgemere Formation metapelites, quartzites, and schists).

The reason why the topsets and foresets indicate opposite flow directions is not clear. The explanation for this difference may provide insight into the nature of the unit we interpret to be topsets, and would have a bearing on the glaciolacustrine versus glaciomarine origin of the delta. As is the case with many of the large marine deltas immediately southeast of this site, the upper portion of the foresets and all of the topset unit may have been removed by regressive wave action or more probably by the late-glacial Saco River. If so, the gravels overlying the foresets may be fluvial stream-terrace deposits. The eastward paleoflow recorded in these gravels supports this hypothesis.

The foreset paleoflow measurements, presence of kettle holes, clast lithology, and proximity of the Saddleback Road Esker and Sand Pond Esker to the north collectively suggest that this delta is an ice-contact delta fed mainly by ice-tunnel drainage from the northeast (Figure 21). Some areas of the delta top exceed 340 ft (103.6 m) in elevation according to the topographic map, and the surveyed topset/foreset

Figure 19. Overview of Stop 7 (Guptill Pit, Baldwin), showing gravelly delta topset unit (left) and sandy foreset beds (right). View to southwest; photo taken in 1993.

Figure 20. Close-up of left part of Figure 19, showing contact (marked by shovel blade) between topset and foreset beds in the Guptill Pit. View to south.

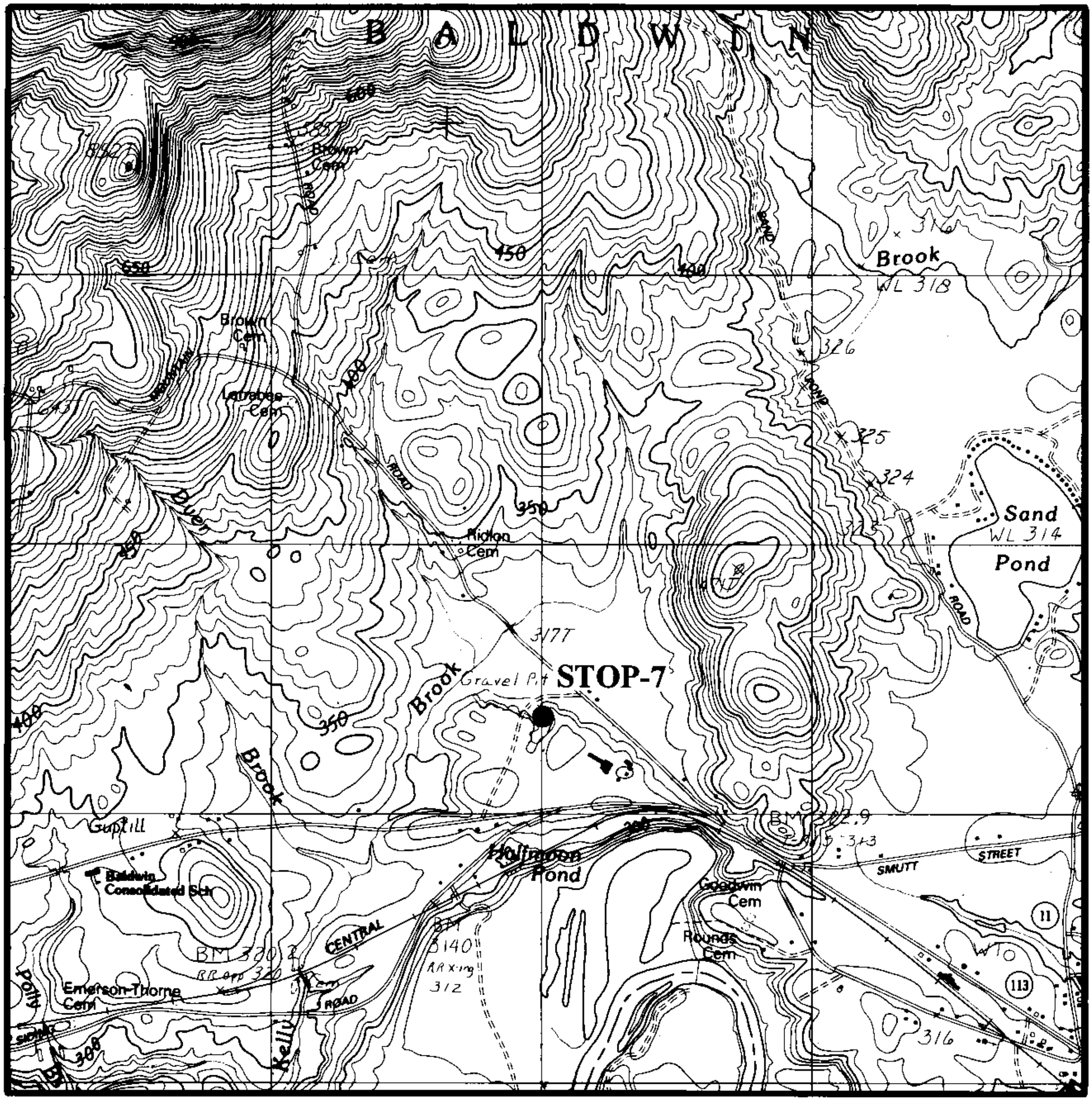


Figure 21. Map showing location of Stop 7 (Guptill Pit, Baldwin) and topography of other glacial deposits in the adjacent Saco River valley.

contact elevation is 332 ft (101.2 m) asl.

It can be argued that the Baldwin Delta is glaciolacustrine. There is no marine clay exposed at this site, although a well log from 4 km down the Saco Valley (Steep Falls village) records 9 m of clay. The latter clay is most likely marine, based on its elevation and association with marine deltas. At Stop 8-A we will see clay exposed upstream along the Saco River. The depositional environment of that clay is equivocal. Just downstream from Stop 7 (in Limington and Standish townships) there is a large delta complex of almost certain glaciomarine origin. Postglacial terracing has modified much of this complex, but the original upper surfaces of the deltas do not extend above 320 ft (97.5 m). The closest reliable topset-foreset contact elevation on a marine delta was surveyed 8 km south of Stop 7, in the North Limington Delta, and is 310 ft (94.5 m) (Thompson and Koteff, unpublished data). However, if we allow for postglacial tilt of up to 0.85 m/km between the two sites (Koteff and others, 1993), it is still possible that the Baldwin Delta is a marine deposit.

We believe the Baldwin Delta was formed in a glaciomarine environment on the basis of scanty evidence. If the topset/foreset contact elevation measured in this gravel pit represents the coeval water level, then the elevation is in close agreement with an extrapolation of the marine limit based on sites to the east and southeast (Thompson and others, 1989; Thompson and Koteff, unpublished data). Although it would be possible (but difficult) to have a sufficiently large ice dam to impound a glacial lake in the Saco Valley south of the Guptill Pit, we have found no evidence of this. (We believe the dam responsible for the lake deposits seen on Day 2 was farther upstream.)

Question:
Is this delta lacustrine or marine?

[End of Day 1 road log]

ROAD LOG FOR DAY 2

Ossipee and Saco River Valleys

0.00	0.00	Start at jct. of Rtes. 5 (N) and 25, on E side of Cornish village. Go E on Rte. 25.
0.70	0.70	Turn R into pit.

STOP 8: CORNISH GRAVEL PIT (Cornish Quadrangle)

Leader: Robert Newton

In the Cornish area, the deposits in the Ossipee Valley are quite different from those within the Saco Valley. The Ossipee Valley deposits have a gentler slope (1.8 m/km) and are predominantly outwash overlying lacustrine sediments, while the Saco Valley sediments are dominated by steeper (3.7 m/km), kettled, ice-contact deposits. After the ice retreated north of the east-west trending Ossipee Valley, this valley acted as a trough which collected south-flowing meltwater and diverted it eastward toward the Saco Valley. The fact that these outwash deposits merge with kettled ice-contact deposits in the Saco Valley (Figure 22) suggests that the Ossipee Valley was ice-free before the lower portion of the Saco Valley. This probably occurred as a result of the blocking of ice flow by the higher hills west and north of Cornish.

The geomorphology in the area of confluence of the Ossipee and Saco rivers is complex. The kettled deposits in the Saco Valley are cut by numerous meltwater stream channels (Figure 23). Some of these incised channels are themselves kettled, suggesting rapid deposition followed immediately by downcutting. The pattern of these channels suggests that they were part of a braided system. It seems that the braided stream was initially responsible for the deposition of the kettled deposits, and that before the buried ice blocks had a chance to completely melt there was a relative decrease in base level which caused these streams to become incised. The secondary channels were then abandoned, leaving only the current channel of the Saco River as the actively eroding channel. The deposition of the sediments, incision of the channels, and their subsequent abandonment all occurred before the buried ice blocks finally melted to form the kettle holes within the incised channels.

It is possible that the incision of the channels was caused not by a relative change in base level (uplift) but instead by a loss in sediment load. This loss would have occurred as a result of the formation of glacial lakes in the upper Saco Valley above Great Falls. The lakes could have removed sediment without significantly affecting discharge, thus causing a period of downstream erosion.

Two of the channels south of Stop 8 are cut into till high on the sides of the valley. These channels were probably cut at an early stage by meltwater escaping eastward out of the Ossipee Valley.

Stop 8 is located in a coarse gravel deposit presumably deposited by meltwater escaping out of the Ossipee Valley. The gravel layers dip eastward and are composed of lithologies consistent with a westward source. The lower sand surface is kettled and is presumably part of the Saco River sequence.

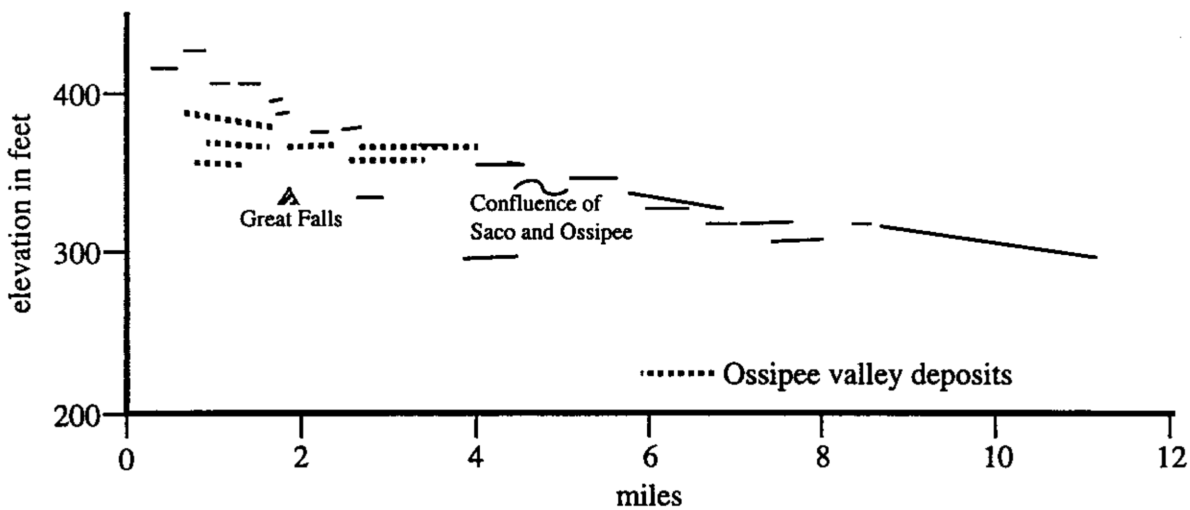


Figure 22. Profile of terrace elevations in the Saco and Ossipee River valleys. Deposits in the Ossipee Valley are primarily outwash, while those in the Saco Valley are mainly ice-contact features.

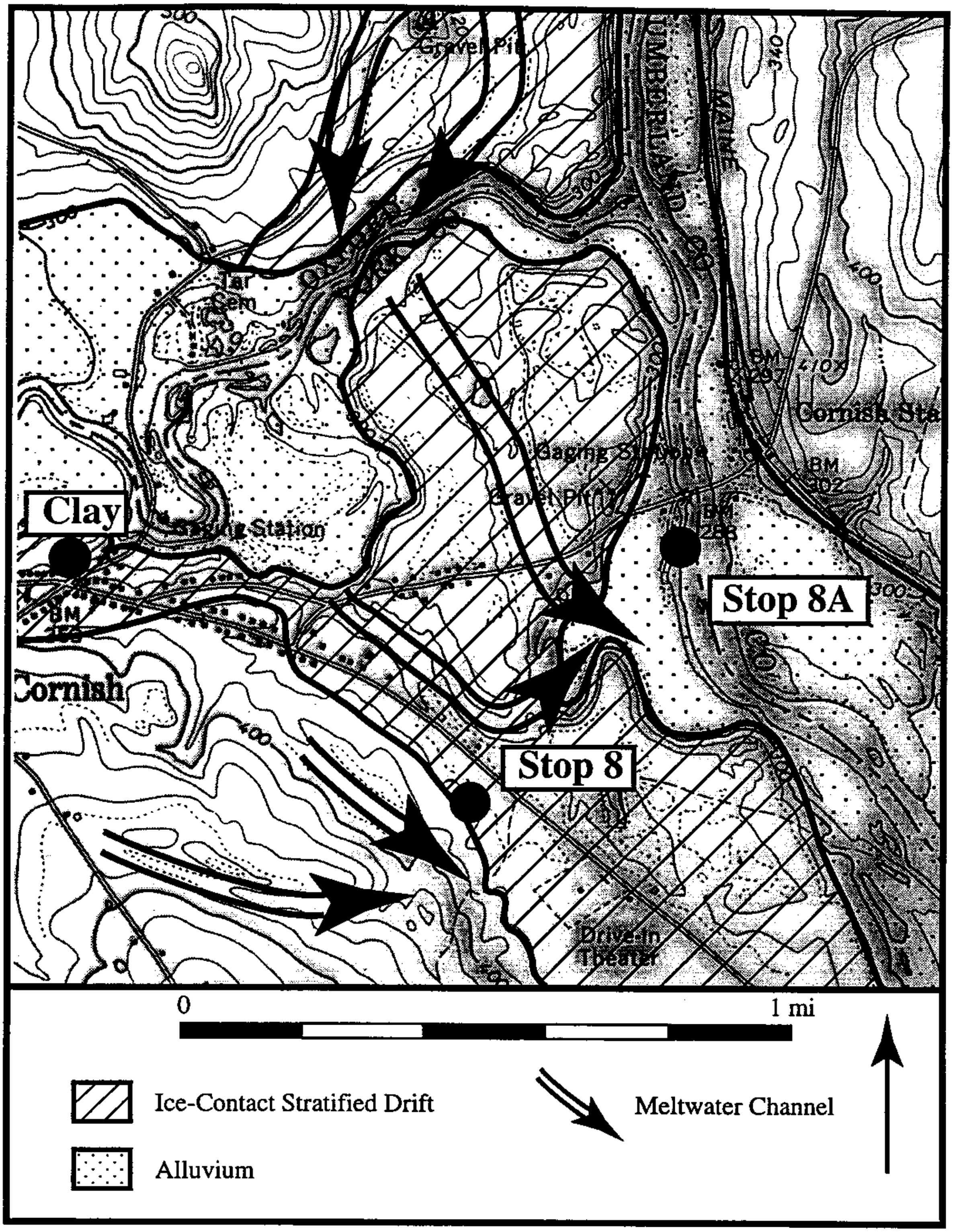


Figure 23. Surficial geology in the area of confluence of the Saco and Ossipee Rivers in Cornish, Maine. Stop 8 is a small pit containing coarse gravel beds dipping eastward. Stop 8-A is a stream cut exposing rhythmically bedded sandy silts and clays. Holland's (1986) rhythmite outcrop is labelled "clay".

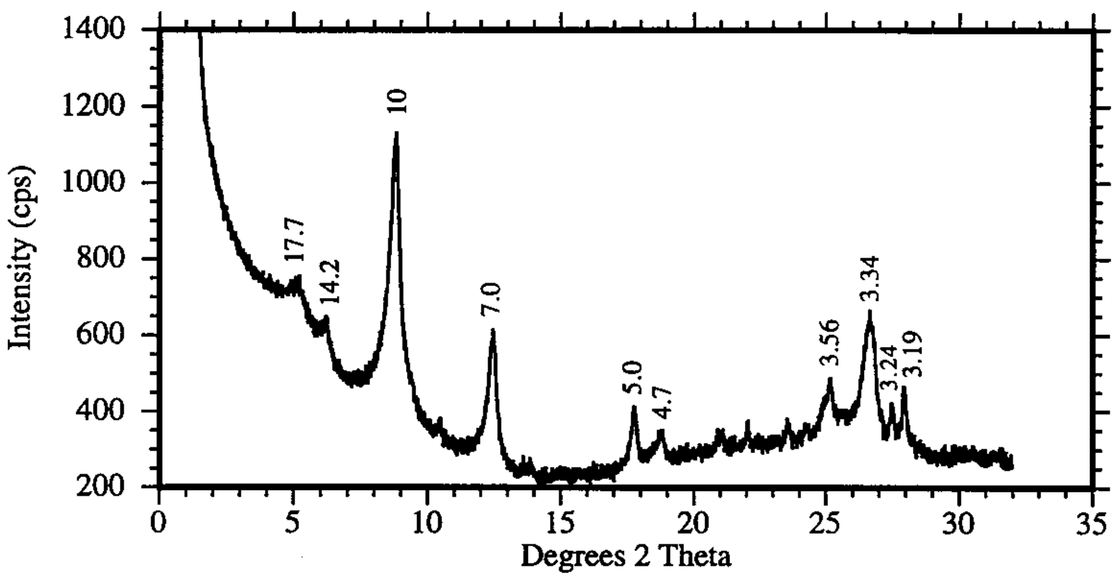


Figure 24. X-ray diffractogram of an oriented <1 μm sample of clay from stop 8-A. The sample has been treated with ethylene glycol in order to identify expandable minerals. The 17.7 Å peak is from smectite or an expandable vermiculite. Other phases present include chlorite, illite and kaolinite.

Questions:
1. Were the rhythmites exposed at this site deposited in the same lake as those in the village of Cornish?
2. How extensive was this lake?
3. Where was the dam for the lake?
4. Could these deposits be estuarine?

Turn around and head back toward Cornish village.

2.40	0.60	Keep R on Rte. 5.
2.70	0.30	Turn R onto Rte. 25 in Cornish and go W through town.
6.10	3.40	Turn L into pit.

STOP 9: WEDGWOOD BROOK ICE-CONTACT DEPOSIT (Cornish Quadrangle)

Leader: Robert Newton

Driving westward from Stop 8 we have traveled across the terraced outwash formed by eastward flowing meltwater streams in the Ossipee Valley. The presence of rhythmically bedded lake deposits revealed in deep well logs and exposed in Cornish suggest that these deposits are deltaic. As the lake in the Ossipee Valley was completely filled by prograding deltas, a smaller lake was left in the valley of the northward-flowing Little River. The level of this lake was probably controlled by the level of the delta in the vicinity of the Cornish Fairgrounds, and this lake did not get completely filled in.

Stop 9 is located in an ice-contact deposit on the south side of the Ossipee valley (Figure 25). These are early deposits formed by meltwater streams flowing south and southeastward parallel to the direction of ice flow. They are typical of deposits found in similar topographic positions across the area. Generally these deposits formed where meltwater streams flowed up and out of a main east-west valley. The deposits include esker segments which appear to be part of larger esker systems that indicate flow in directions independent of the surrounding topography. The deposits are kettled and include large boulders and poorly sorted sediments which are generally associated with heads of outwash.

The Wedgwood Brook ice-contact deposit extends up to an elevation of 500 to 520 ft (152-158 m). There is a bedrock "island" in the center of the deposit which is completely surrounded by stratified drift. Numerous esker segments cut across the feature, with some of the larger ones on the south side being over 15 m high. After this deposit formed it appears to have acted as a dam for the northward flowing Wedgwood Brook, creating a small lake in the area of Spruce Pond. The outlet for this lake appears to be a bedrock spillway between Spruce Pond and Long Pond.

Questions:
1. Why was fluvial deposition concentrated at this location?
2. Does this site mark a boundary between active and stagnant ice?

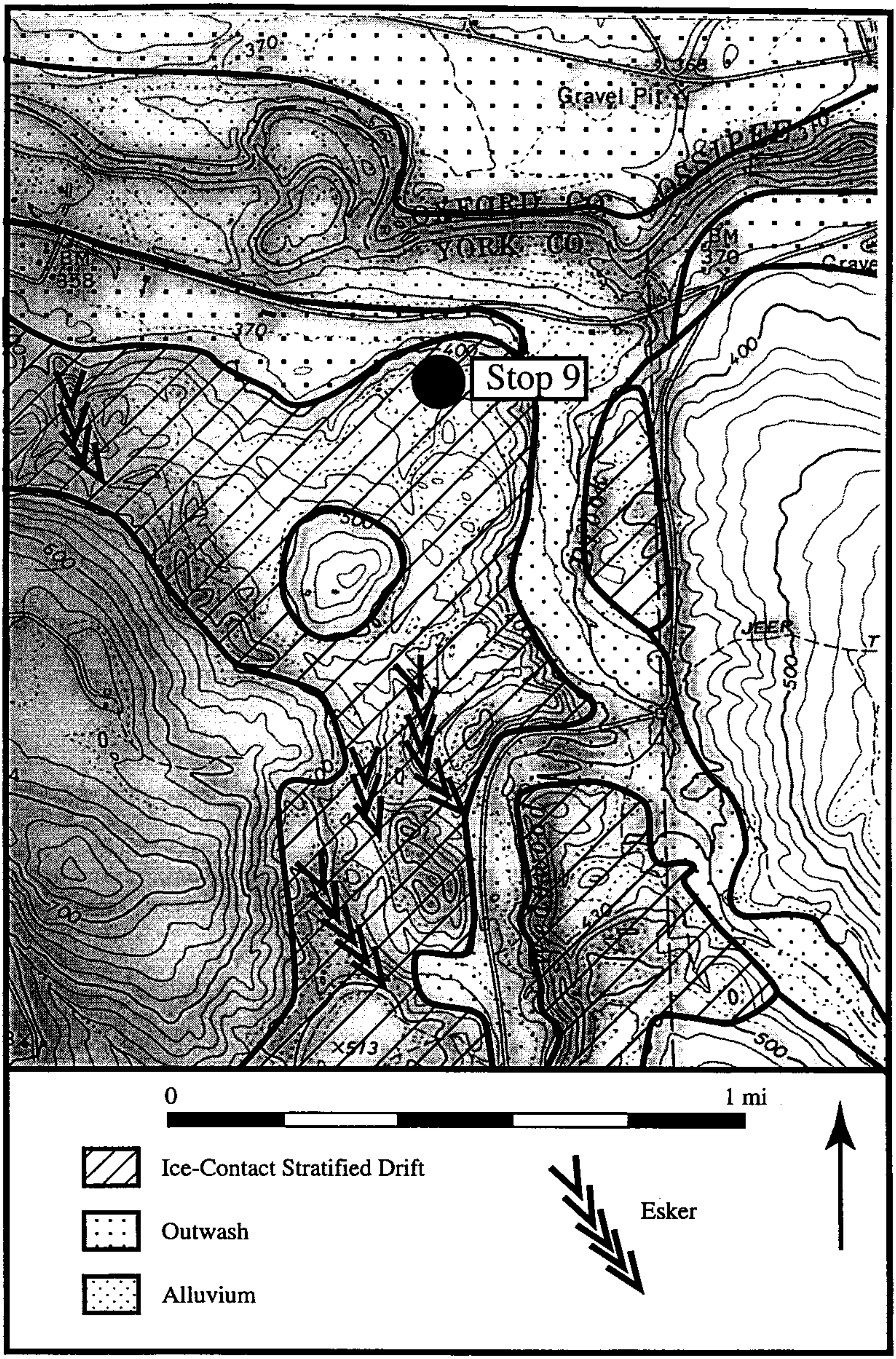


Figure 25. Surficial geology in the area of Stop 9. Stop 9 is located in an ice-contact complex which was formed by meltwater flowing southward.

Continue W on Rte. 25.

7.40	1.30	Just before bridge in Kezar Falls village, turn L onto Elm St., which soon turns to S.
10.80	3.40	Turn R (W) onto Mudgett Rd.
11.70	0.90	Park on R and walk N on woods road about 2300 ft to area of small ridges.

STOP 9-A: GREAT BROOK ROGEN MORAINE (Kezar Falls Quadrangle)

Leader: P. Thompson Davis

The curvilinear ridges in the Great Brook area, composed of cobbly diamict with a sandy matrix (Figure 26), were first described in detail by Holland (1986). Similar ridges are found elsewhere in the Kezar Falls and adjacent quadrangles, but only in valley bottoms that trend generally north-south to S 30° E, which is the direction of last glacial flow as inferred from bedrock striae and stoss-lee forms. The ridge segments in the Great Brook area lie at elevations between 400 and 450 ft (120-140 m) and are generally 10 to 30 ft (3-9 m) high, 50 to 150 ft (15-45 m) wide, and less than 500 ft (150 m) long. Similar, but larger, ridges have been described in the Millinocket Lake basin in west-central Maine at altitudes below 640 ft (195 m) and only overlying phaneritic plutonic rock (Caldwell and others, 1985; Weddle and others, 1994). The ridges are generally hummocky, segmented, and sub-parallel to one another, and usually occur in clusters or "fields," with their long axes oriented transverse to long axes of valleys, which drain both north and south. These ridges are mapped as "ribbed moraine" on the Surficial Geologic Map of Maine (Thompson and Borns, 1985). Holland (1986) noted that length-to-width ratios between 4:1 and 10:1 make the ribbed moraines in the Great Brook area stubbier than the DeGeer moraines in coastal Maine described by Smith (1982, 1985).

Similar ridges in the Millinocket Lake basin south of Mt. Katahdin have been called Rogen moraine (no 's' at end) by Caldwell and others (1985), after moraines of similar appearance near Lake Rogen, Sweden, described and named by Jan Lundqvist (1969, 1981, 1989). Lundqvist (1989) discussed four different hypotheses for the origin of Rogen moraine: (1) deposition as subglacial moraines formed under thick ice away from the ice front, perhaps in a transition zone between warm- and cold-based ice under compression; (2) deposition as marginal moraines, forming a complex of end moraines; (3) formation by tectonic processes within active ice, such as folding or thrusting of debris-rich layers against obstacles to glacier flow, followed by melt out; and (4) filling of either open crevasses by supraglacial debris or basal cavities by subglacial debris. Sugden and John (1976) and Bouchard (1989) favor shearing or thrusting processes up-glacier from the snout for the origin of Rogen moraine. Fisher and Shaw (1992) favor subglacial origin for Rogen moraine, classifying the ridges as an end-member of transitional subglacial streamlined features with drumlins being the opposite end-member. Although Lundqvist (1989) uses the terms "Rogen" and "ribbed moraine" interchangeably, he limits usage to a specific geographic region of Sweden.

West of the Millinocket area in west-central Maine, Caldwell and others (1985) described Rogen moraine that are cut by an esker, and thus interpreted the moraine to be older than the esker and subglacial in origin. Weddle and others (1994) agreed that many of the moraines in the Millinocket area are subglacial in origin, but pointed out that they do not always grade into drumlins and are not everywhere cut

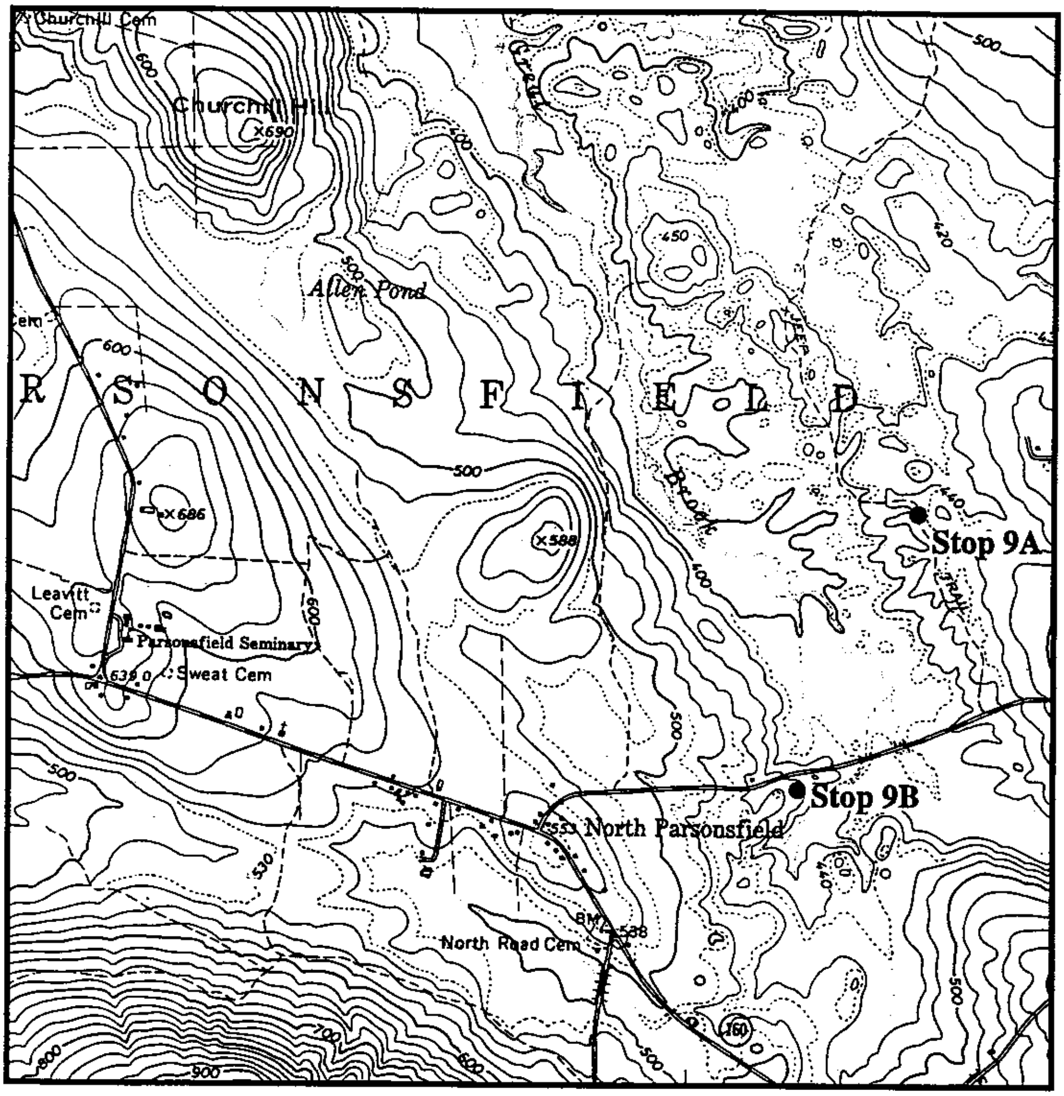


Figure 26. Part of Kezar Falls 7.5-minute quadrangle, showing field of curvilinear ridges in Great Brook valley (Stop 9-A) and a small mound of ice-contact stratified drift on the valley side (Stop 9-B).

by eskers; they note instead that eskers in many areas broaden and diffuse into hummocky ice-contact topography. In northern Quebec, Bouchard (1989) identified ridges that he interpreted to be Rogen moraine formed subglacially in lee-side basins by stacking of debris-rich ice against obstacles as the front part of the glacier stagnated. A similar mechanism was proposed for the origin of ribbed moraine in the Millinocket area, which is down-flowline from Mt. Katahdin, which in effect could have cut off the sediment supply during late phases of deglaciation (Bjorn Andersen, in report to Maine Geological Survey, 1992, in Weddle and others, 1994). Weddle and others (1994) proposed that eskers and adjacent moraines are essentially contemporaneous and mark ice-marginal positions, and preferred the terms "ribbed" or "transverse" moraine, as either is less generic than Rogen.

A till fabric was obtained in an exposure for a logging road through a small ridge oriented about N 20° E on the 420-ft contour, just south of the "D" in "PARSONSFIELD" on Figure 26. This ridge segment, which is about 0.1 mile (150 m) long, is oriented more obliquely to the general N 40° W trend of Great Brook valley than other ridge segments in this area. One hundred elongate clasts (2 to 10 cm long axes; greater than 1.5:1 long:intermediate axis ratio) were measured for trend and plunge. A strong fabric was found, with an average trend about N 65° W and an average plunge about 8° toward S 65° E (Figure 27). These directions are roughly perpendicular to the orientation of the ridge. Such a strong fabric is consistent with a subglacial origin by meltout or lodgement. Holland (1986) noted that stratified sediments generally overlap both distal and proximal faces of the ridges, and suggested that these sediments could have been deposited under thick ice during the early phases of deglaciation. More fabric analyses are planned for adjacent ridge segments in the valley.

Questions:

1. Are these ridges indeed Rogen moraine?
2. Were these ridges formed subglacially or at the glacier margin?
3. What is the relationship of these ridges to channel forms on the valley side?
4. What is the relationship of these ridges to nearby glaciofluvial deposits?

Return to Mudgett Rd. and continue west.

12.10 0.40 Turn L into pit.

STOP 9-B: MUDGETT ROAD DEPOSIT - Kame delta or ice-channel filling? (Kezar Falls Quadrangle)

Leader: P. Thompson Davis

This active pit is cut into a 15-ft (4.5 m) high, 300-ft (100 m) diameter mound (Figure 26) and revealed excellent exposures of well-stratified and well-sorted fine to medium-grained sands and pebbly sands along the north-facing wall in October, 1994. The coarser sands showed tabular sets of graded and reverse-graded beds, with ripple cross stratification and laminated drapes indicating a generally S 40° E flow direction. The exposure is capped by 1.5 ft (0.5 m) of fine sand and silt (loess?). Holland (written communication, 1985) reported that a few sharply-defined normal faults were also present, but did not note

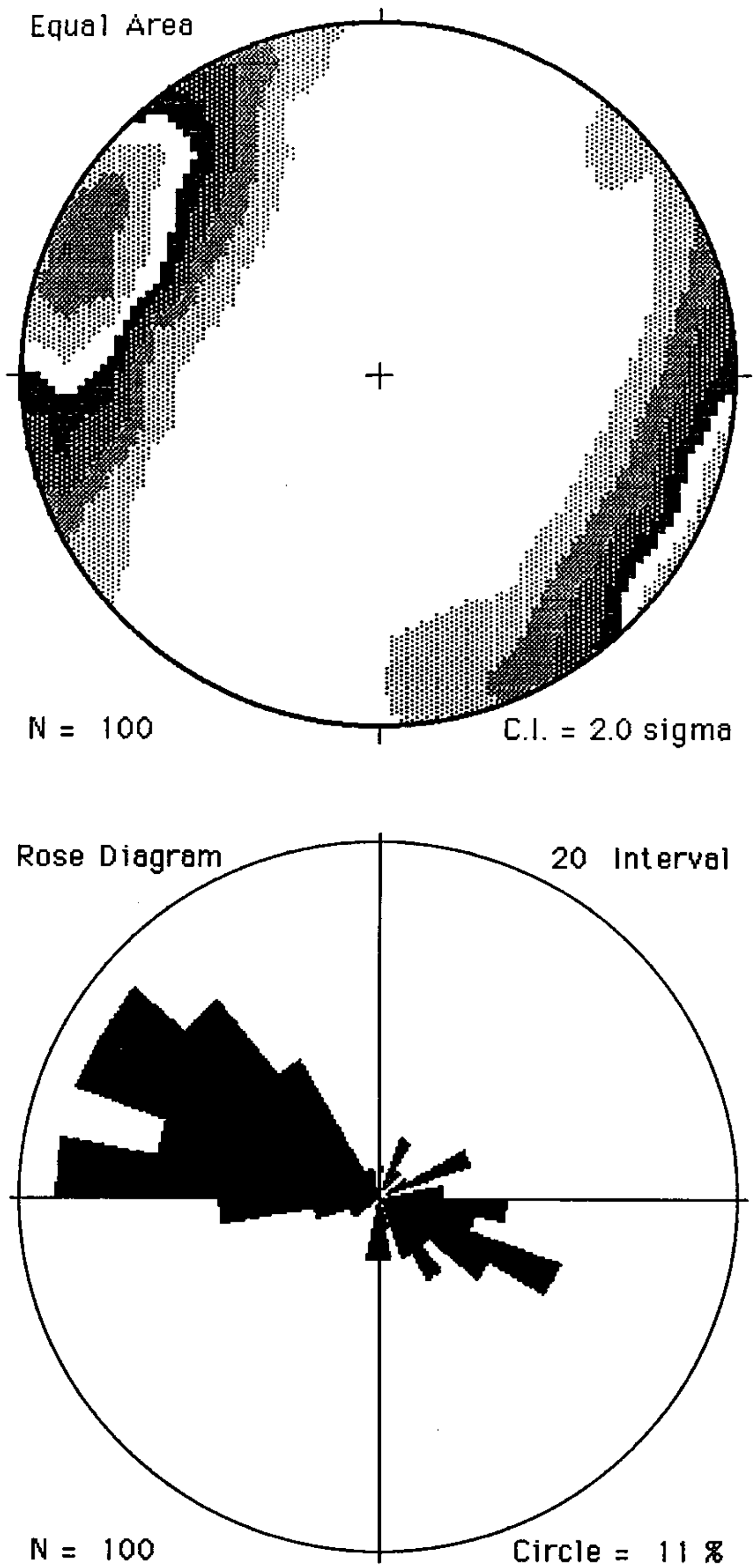


Figure 27. Equal-area and rose diagrams for till fabric analysis from curvilinear ridge in Great Brook valley near Stop 9-A.

whether these faults extended through a loess cap. Faulting was not visible in October, 1994. Holland (written communication, 1985) suggested that the deposit could be an ice-channel filling, but was probably not a true tunnel deposit. The hummocky topography and the location of the deposit on the southwest side of the upper part of the north-draining Great Brook valley could also be suggestive of a kame or kame-delta, deposited during the waning phases of deglaciation of the area. Less well-stratified very coarse sand, gravel, and till, with boulders up to 1.5 m in diameter, are exposed in the east-facing pit wall. The till exposed here was described as ice disintegration moraine by Holland (1986).

Questions:

1. What is the origin of this deposit?
2. What is the relationship of the stratified sediments exposed in the middle of the pit to the non-stratified deposits exposed on the west side of the pit?
3. What is the relationship of this deposit to the curvilinear ridges in the valley to the north?

Continue W on Mudgett Rd.

12.60	0.50	Turn R onto Rte. 160 at North Parsonsfield.
13.50	0.90	Turn R (N), staying on Rte. 160.
15.70	2.20	Cross Ossipee River (note covered bridge on R).
16.00	0.30	Turn R (E) on Rte. 25.
16.20	0.20	Turn L (N) onto Old County Rd. in Porter village.
17.70	1.50	Turn L at stop sign, onto Old Meetinghouse Rd.
18.00	0.30	Keep L at jct., going NW on Colcord Pond Rd.
20.50	2.50	Bear L at fork.
20.60	0.10	Bear R onto Kennard Hill Rd. (becomes Porter Rd. farther N).
23.70	3.10	Keep L at fork.
24.80	1.10	Turn L, then quickly to R, onto old woods road leading N to pit.

STOP 10: SHEPARDS RIVER - COLE BROOK CLIMBING ESKER (Brownfield Quadrangle)

Leader: P. Thompson Davis

Stone (1899, p. 257-258) wrote "The number and height of the hills which the gravels of this region cross are remarkable. Nowhere else in Maine is there anything equal to them. In Brownfield, Porter, and Hiram the glacial rivers flowed up and down over these hills 200 or more feet higher than the valleys to the north of them, and in Parsonfield and Cornish they crossed several more ... These branching series often reject valleys of favorable slopes to climb hills, and are therefore difficult to map. Delta branches are liable at any point to diverge from the series one is exploring, and constant watchfulness is required." The Shepards River - Cole Brook esker system (shown in part on Figure 28) remains more readily examined on map than in section.

Most exposures in the Shepards River - Cole Brook esker system, which extends over 7.5 mi (12 km) and nearly the length of the west side of the Brownfield Quadrangle, are extremely slumped and no longer

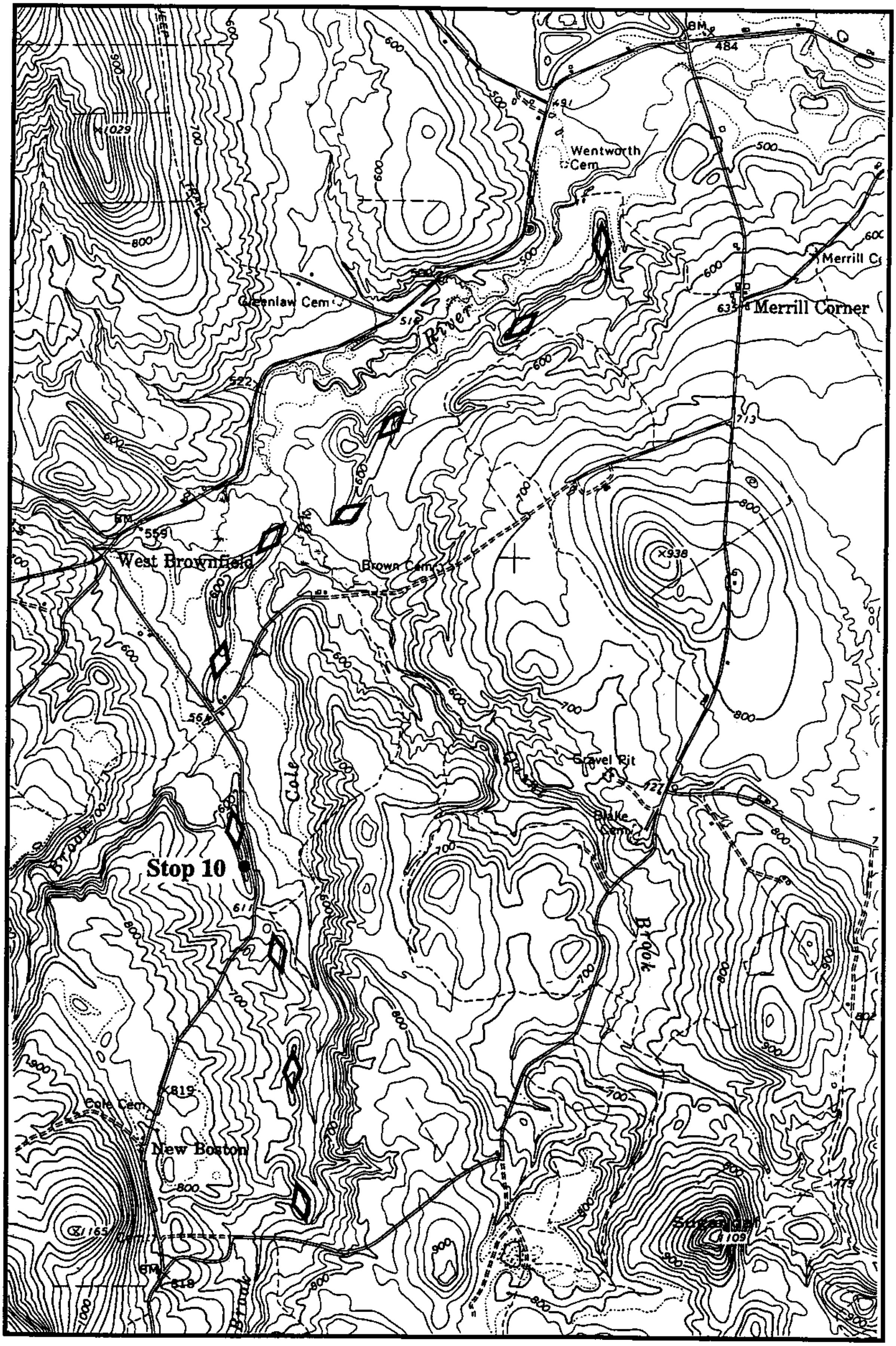


Figure 28. Part of Brownfield 7.5-minute quadrangle, showing esker that climbs over 330 ft (100 m) in elevation from north to south along Shepards River and Cole Brook (Stop 10).

useful for study. However, in 1985, Holland and others (Carl Koteff, 1995, oral communication) examined exposures that provided unequivocal paleocurrent data that indicated flow from north to south. Also, this esker climbs over 330 ft (100 m) in elevation, from near 500 ft (150 m) to the north to over 830 ft (250 m) to the south, where a bedrock topographic divide marks the lowest point in the upper Cole Brook drainage. Thus, the esker climbs about 44 feet per mile (about 8.3 m/km). Holland (1986) noted a lack of collapse features within the core of the esker, which he suggested argued against superposition of the esker on the bed from an englacial position. He favored deposition from a subglacial position in a confined tunnel. Thus, Holland (1986) suggested that the esker was formed during an early phase of deglaciation when ice was still thick, whereas other ice-contact stratified deposits in the valley were formed later when ice had thinned. This two-stage model for deposition of ice-contact stratified drift is similar to one proposed by Goldthwait and Mickelson (1982) for deglaciation of the White Mountains in New Hampshire.

Elsewhere in Maine and New Hampshire, Shreve (1985a, b) and Boothroyd (1985) argued that eskers represent through-flowing systems from end-to-end that were deposited simultaneously within continuously open tunnels. Thus, Shreve (1985a, b) claimed that the Katahdin esker system in central Maine could be used to reconstruct hydraulic conditions and ice profiles for the ice sheet, hence ice thickness for any location along a flowline. However, Weddle and others (1994) countered that the Katahdin esker system consists of several distinct segments separated by fans or deltas, and thus was not deposited concurrently, but rather was deposited sequentially in segments by retreating active ice. Thus, this latter view is consistent with the deposition of morphosequences and end moraines in New England as described by Koteff and Pessl (1981).

The Shepards River - Cole Brook esker is not nearly as large or long as the esker systems in central and eastern Maine and does not appear to exhibit fans or deltas at segment ends. The segments in the Shepards River - Cole Brook esker are probably the result of differential preservation on a topographically variable bedrock floor. With limited exposures, the question concerning whether this esker was formed as one contemporaneous feature or deposited sequentially as individual segments is open to discussion. However, the steep longitudinal gradient of this esker might argue for continuous flow and thick ice to provide the necessary hydrostatic head to climb southward.

Although the large pit at this site is severely slumped and not very active, the pit extends into the core of the esker and allows us the opportunity to examine coarse sand and gravel bedding typical of esker deposits in Maine.

Questions:

1. Can we see any flow indicators in the present limited exposures in the pit?
2. How thick was the surrounding ice when this feature was deposited?
3. What is the temporal relationship between this deposit and those seen at Stops 9-A and 9-B?
4. Was the esker formed as one contemporaneous feature, or was it deposited sequentially as individual segments?

Return to road.

24.90	0.10	Turn L on Porter Rd. and continue N.
25.95	1.05	Turn R onto Eaton Center Rd. at West Brownfield.
30.30	4.35	Turn R (SE) onto Rte. 160 at fork in Brownfield village.
30.50	0.20	Park on side of road. Note terraced delta along road and in cemetery to E of road.

STOP 10-A: BROWNFIELD DELTA - Terraced delta deposited in Lake Pigwacket (Brownfield Quadrangle)

Leader: P. Thompson Davis

The Pine Grove Cemetery offers a fine drive-by "stop" to view at least two, and perhaps three, terraces on a large delta deposited into glacial Lake Pigwacket. The next two stops will allow examination of exposures of deltaic and lake bottom sediments. The delta extends almost 2 mi (about 3 km) southwest from the Saco River between Shepards River and Burnt Meadow Brook (Figure 29). The delta generally lies above the 400 ft (122 m) contour. Terraces at the cemetery occur between about 420 and 435 ft (128 and 133 m), whereas a terrace along the Saco River to the northeast lies at about 380 ft (116 m), suggesting that meltwater flowed from west to east. Thus, the deposition of the Brownfield delta occurred after the formation of the Shepards River - Cole Brook esker system.

Continue SE on Rte. 160.

31.10	0.60	Bear R at jct., staying on Rte. 160.
31.40	0.30	Turn R into pit.

STOP 11: BROWNFIELD TOWN PIT - Lake Pigwacket delta (Brownfield Quadrangle)

Leader: P. Thompson Davis

This is an active pit for the town of Brownfield (Figure 29). During October, 1994, excellent exposures at the south end of the pit exhibited about 20 ft (6 m) of foreset beds consisting of coarse sand and gravel with dips of about 20° in a generally southeastward direction. Approximately 3 ft (1 m) of topset beds composed of pebble-cobble gravel were exposed in other parts of the pit. The elevation of the topset/foreset contact was about 420 ft (128 m), but this contact was not clearly exposed. During early April, 1995, all exposures were very slumped, thus the foreset beds were barely visible. Nevertheless, the pit is believed to exhibit sediments of an upper delta facies, following the nomenclature of Ashley and others (1982, 1985), that were deposited into glacial Lake Pigwacket.

Questions:

1. What is the origin of this deposit?
2. What is the elevation of the topset/foreset contact, if visible?
3. What is the relationship of this deposit to the terraces exposed at Stop 10-A?

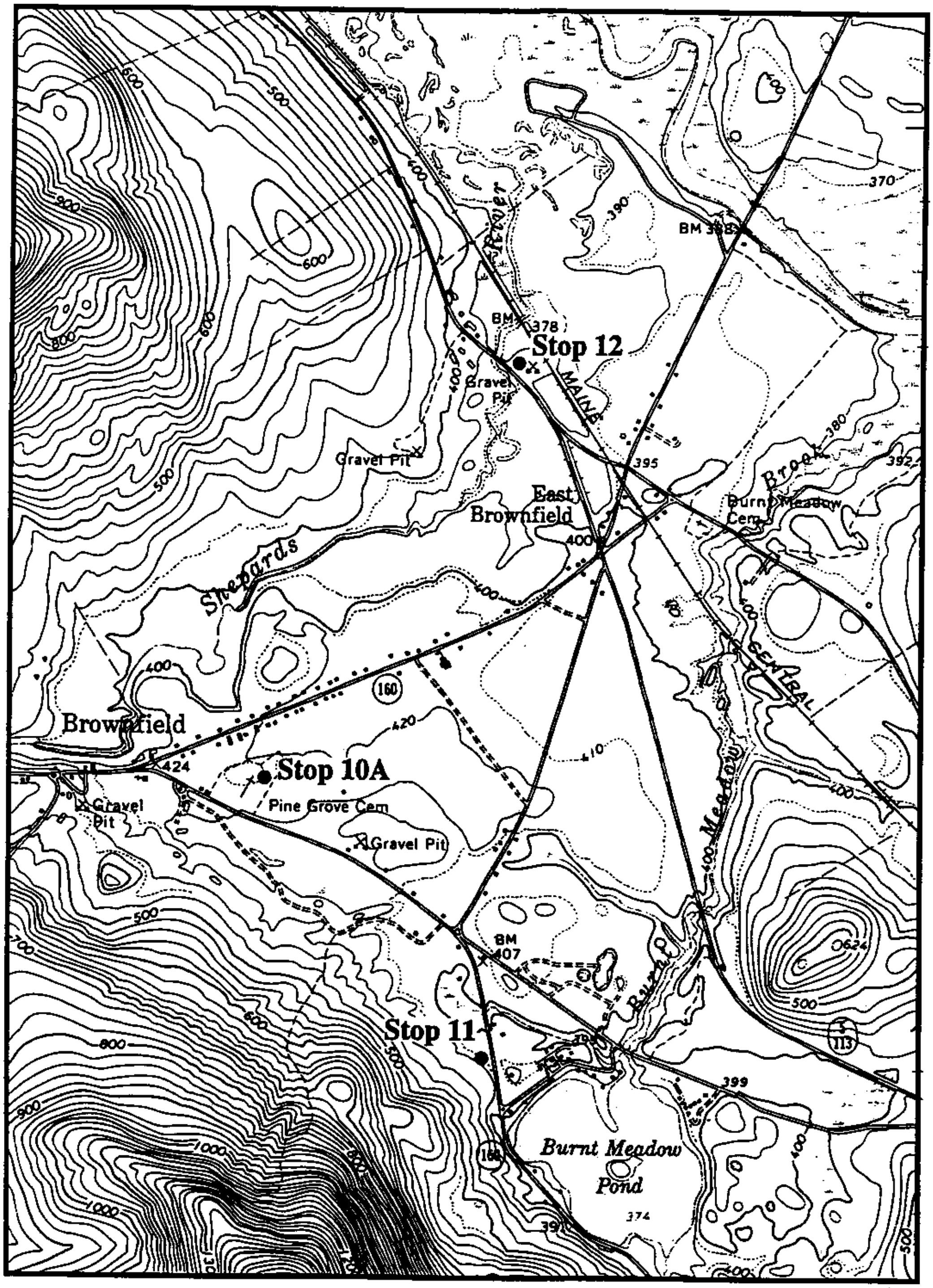


Figure 29. Part of Brownfield 7.5-minute quadrangle, showing deltaic deposits related to glacial Lake Pigwacket (Stops 10-A, 11, and 12).

Leave pit and head back toward Brownfield on Rte. 160.

31.80 0.40 Turn R (<u>not</u> hard right) and drive NE across delta.
32.60 0.80 Turn L onto Rtes. 5/113.
33.10 0.50 Turn R into small pit next to garage.

STOP 12: BARRETT PIT - Lake Pigwacket sediments (Brownfield Quadrangle)

Leader: P. Thompson Davis

This pit was recently backfilled for construction of a large garage-warehouse. However, excellent exposures of mid-delta and lower delta foresets (Ashley and others, 1982, 1985) are exhibited in the remaining 10-ft (3-m) high exposure on the north side of the pit. The elevation of the top of the section is about 380 ft (116 m) (Figure 29). Ronald Barrett has graciously offered to deepen the exposure for further examination during our trip, and wishes to make the site semi-permanent for future observations.

The exposure along the north wall of the pit exhibits all three types of climbing ripple-drift described by Ashley and others (1982, 1985) as indicative of mid-delta and lower delta facies (Figure 30). Type A ripples show stoss-side laminae that have been eroded by rapidly migrating ripples with low vertical aggradation, whereas Type B ripples have preserved stoss-lee laminae resulting from slowly migrating ripples with high vertical aggradation during a waning flow (Figure 31). Draped laminations form when ripple migration ceases and finer-grained sediments fall from suspension.

The sediments at this locality consist mostly of fine sand and silt, although coarser sand and clay layers are also present. The variety of ripple types suggests variable meltwater discharge into a lake. Flume experiments by Ashley and others (1982) indicate that such ripple sequences may form very rapidly (i.e. in a few hours), thus many flow events are possible within one season. The direction of ripple climb is about N 45° E to N 55° E, so flow into glacial Lake Pigwacket at this site was from the southwest. Loading and dewatering structures are also common here.

Questions:

1. Are the sedimentary features in the Barrett Pit actually climbing ripple-drift cross strata?
2. Are these sediments consistent with glacial meltwater deposition into a shallow lake?
3. What is the relationship of this deposit to the esker at Stop 10, the terraces at Stop 10-A, and the upper delta sediments at Stop 11?

Leave pit and go S on Rtes. 5/113.

40.05 6.95 Turn L onto Rte. 117 (just past Saco River bridge), and drive straight through Hiram village.
41.65 1.60 Park carefully on side of Rte. 117 **(watch for traffic!)** and walk uphill in woods E of road.

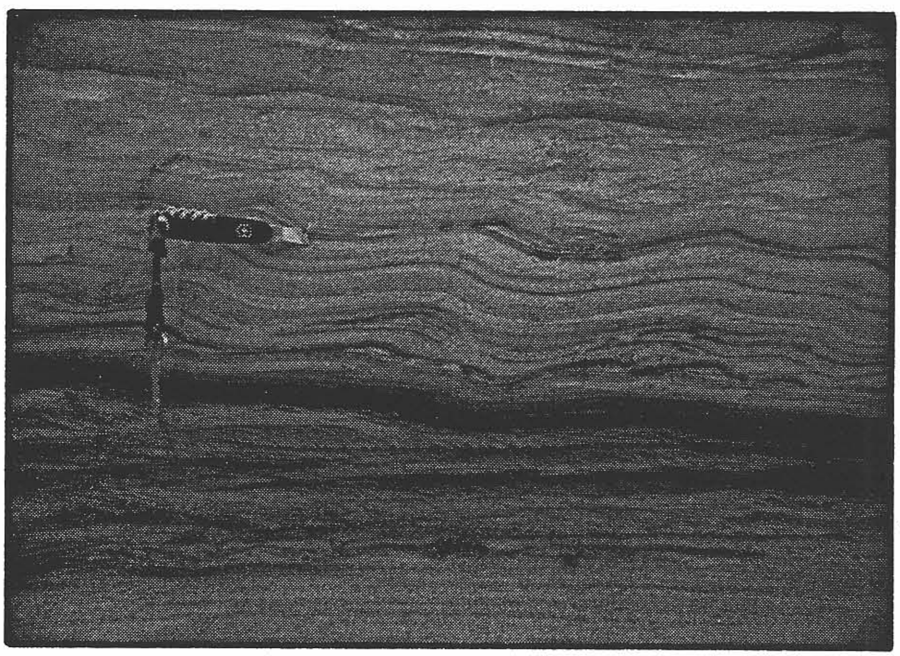


Figure 30. Photographs of lake-bottom sediments exposed at the Barrett Pit that exhibit a variety of ripple-drift cross stratification (Stop 12).

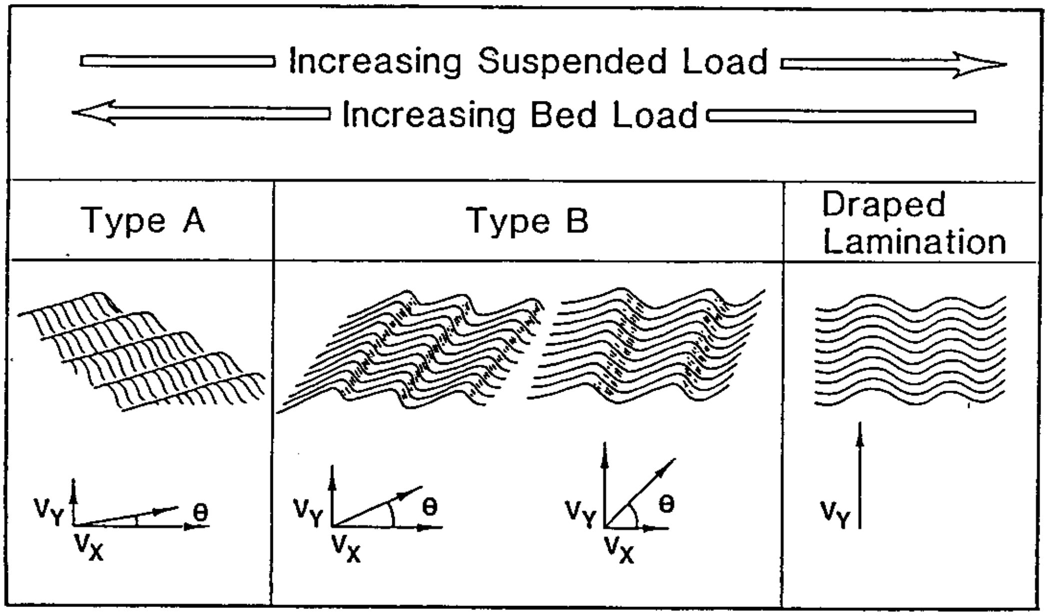


Figure 31. Diagram from Ashley and others (1985, p. 189), showing three types of climbing ripple-drift sequences. The ripples climb at an angle, θ, whose tangent is the mean aggradation rate (V_y) divided by the downstream migration rate (V_x) (after Ashley and others, 1982). All three types of ripple drift were observed in the Barrett Pit.

STOP 12-A: HIRAM MORAINES (Hiram Quadrangle)

Leader: Woodrow Thompson

Recessional ice-margin positions in the Hiram Quadrangle are indicated by hillside meltwater channels, and by various types of ice-contact deposits. Along Route 117 north of Hiram village, there are several bouldery cross-valley ridges. These ridges are tentatively identified as end moraines, based on their morphology, orientation, and surface characteristics. Stop 12-A (Figure 32) is located on a pair of moraine ridges that plug a gap between two hills.

On the east side of the road, there are two very distinct east-west ridges whose crests rise toward the valley wall. The northern ridge is the higher of the two. It is steepest on the north (proximal) side, where there is also a greater concentration of boulders. West of the road, the continuation of the southern ridge is well defined, but the northern ridge becomes more diffuse and hummocky. No bedrock outcrops have been seen on these ridges. Although there are no exposures of the sediments comprising the ridges, the surface boulders suggest that diamict (till) composes at least part of them.

Assuming the ridges are moraines, it should be noted that they are much higher than the Rogen moraine ridges seen at Stop 9-A. Also, these and several other moraines in the southeast part of the Hiram Quadrangle occur as isolated ridges or small groups, and not as clusters of many individuals. However, along with other moraine types in this region (Holland, 1986), they share the characteristic of occurring only in valleys. The size of the Hiram moraines suggests deposition at an active ice margin, but there is no proof of this. There was a late-glacial shift in ice-flow direction over much of interior southwestern Maine (Thompson and Koteff, 1995), resulting in a southward flow phase that is recorded by striations and rarely by glaciotectonic structures. This late, topographically influenced ice activity may have formed the widely scattered moraines at Hiram and elsewhere.

Questions:

1. Why are these deposits concentrated in small valleys, with little or no continuity across adjacent uplands?
2. Are these ridges actually end moraines?
3. Were these deposits formed by active ice?

Return to jct. with Rtes. 5/113 in Hiram.

43.25	1.60	Turn L onto Rtes. 5/113 and drive S.
45.30	2.05	Park and walk into pit on E side of road. Watch for traffic.

STOP 13: GREAT FALLS DELTA - Delta and dam for Lake Pigwacket (Cornish Quadrangle)

Leader: Robert Newton

Stop 13 is located within a delta having a surface elevation of 400 ft (121.9 m) (Figure 33). Similar

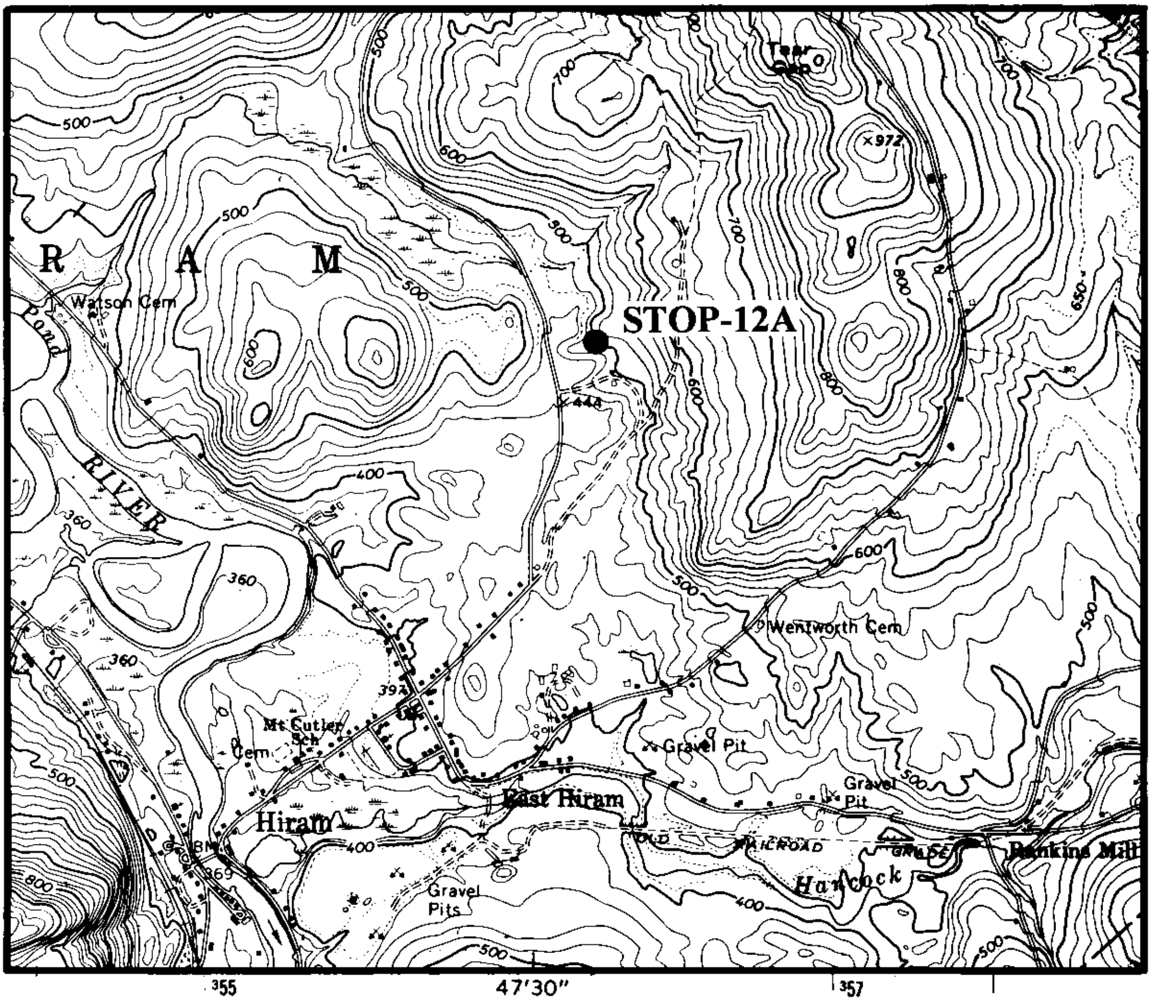


Figure 32. Location map for Stop 12-A (Hiram moraines).

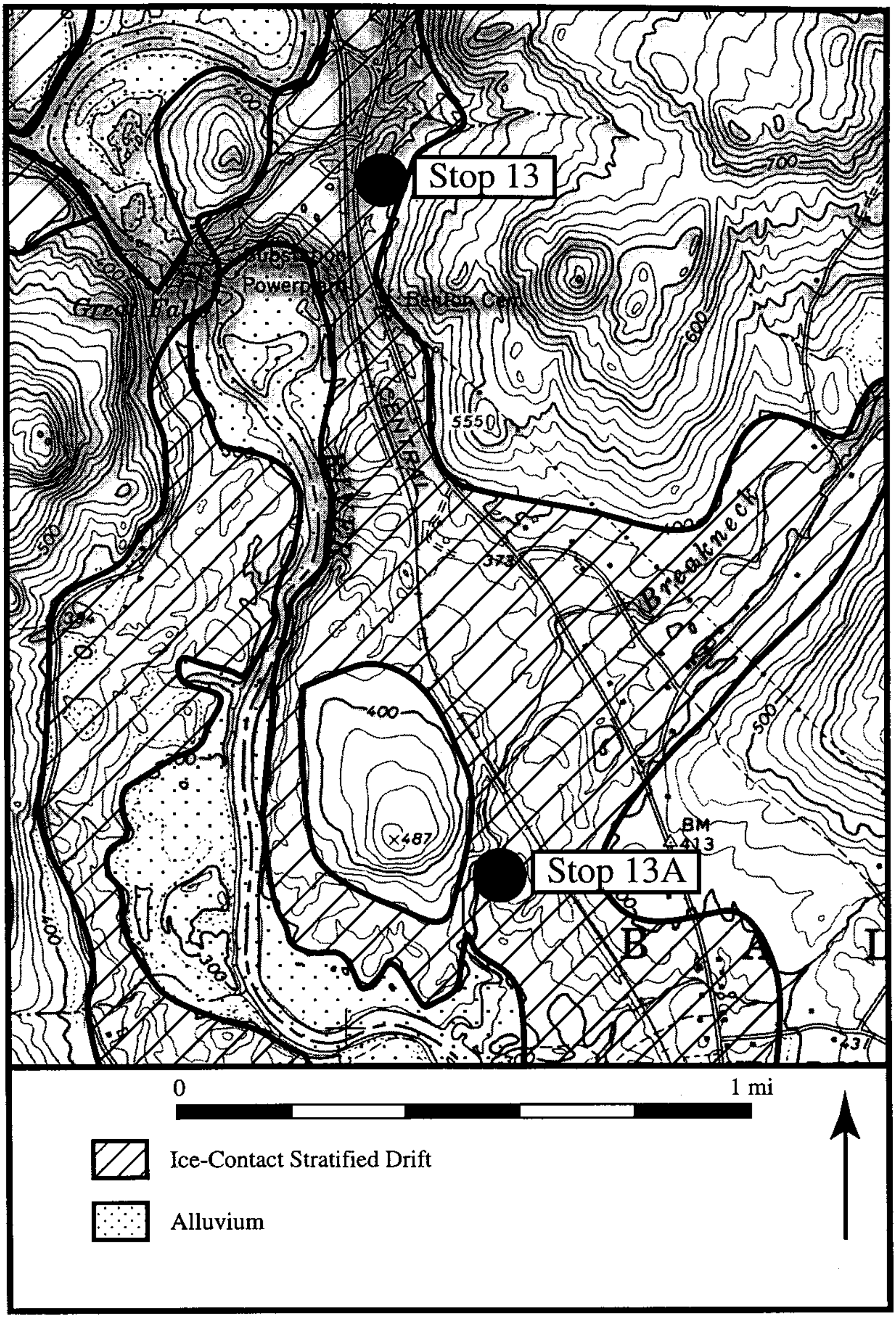


Figure 33. Surficial geology in the vicinity of Stops 13 and 13-A. Stop 13 is located in an ice-contact complex which was formed by meltwater flowing southward.

features are found across the river at elevations of 400 to 410 ft. All are believed to be part of a single delta built into a small ice-marginal lake. Upon further retreat of the ice, the delta acted as a dam for Lake Pigwacket to the north (Thompson, in prep.). The deposits within the Saco River valley north of this point all appear to be graded to the level of these deposits. Erosion of the dam led to the eventual draining of the lake.

The upper part of the pit exposes approximately 2 m of coarse cross-bedded sands and gravels. The dominant current direction is southwest. In the lower areas of the pit there are fine silts and sands with climbing-ripple cross laminations.

South of this site the valley fill is dominated by ice-contact deposits, while north of it, outwash and lake sediments are most abundant. A basic question here is why did the change occur at this location? What is the importance of the bedrock spillway at Great Falls? Did the extensive bedrock outcrops along the west side of the valley form a high-level spillway for Lake Pigwacket? At this point these questions are mainly unresolved.

Continue S on Rtes. 5/113.

46.45 1.15 Turn R and drive into pit complex.

STOP 13-A: MAIETTA PIT (Cornish Quadrangle)

Leader: Robert Newton

Extensive mining at this site has exposed a large section of the ice-contact valley fill typical of this part of the Saco Valley. The deposits here range from till at the bottom of the pit through very coarse cobble gravel to fine lacustrine sands and silts. Much of the sediment is deformed into large folds cut in places by high-angle faults. Some of the sand and gravel beds are actually overturned and have lacustrine sand and silt cutting unconformably across the overturned limb of the fold. In other areas the lacustrine sediments completely fill small kettle holes which are now exposed as "islands" within the pit.

These deposits were formed in part by meltwater streams flowing down the Saco valley and also by meltwater spilling across low divides in the upper part of Breakneck Brook. This suggests that the active ice front was just north of Breakneck Brook at the time these sediments were deposited. The fact that this part of the Saco valley was not flooded by outwash sediments as the ice retreated northward can probably be explained by the trapping of sediment in Lake Pigwacket.

Question:

How does the age of these features compare with the outwash in the Ossipee River valley?

[END OF TRIP]

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APPENDIX

FRIENDS OF THE PLEISTOCENE REUNIONS
1934-1995

Reunion	Leaders	Area
1. 1934	George White / J.W. Goldthwait	Durham to Hanover, NH
2. 1935	Dick Flint	New Haven to Hartford, CT
3. 1936	Kirk Bryan	SE Rhode Island to Cape Cod, MA
4. 1937	J.W. & Dick Goldthwait / Dick Lougee	Hanover to Jefferson, NH
5. 1938	Charlie Denny / Hugh Raup	Black Rock Forest, NY
6. 1939	Paul MacClintock / Meredith Johnson	Northern NJ (drifts)
7. 1940	Kirtley Mather / Dick Goldthwait	Western Cape Cod, MA
8. 1941	John Rich	Catskill Mtns., NY
1942-45	*no meetings during war years*	
9. 1946	Lou Currier / Kirk Bryan	Lowell-Westford area, MA
10. 1947	Earl Apfel	Eastern Finger Lakes, NY
11. 1948	D.F. Putnam / Archie Watt / Roy Deane	Toronto to Georgian Bay, ONT
12. 1949	Paul MacClintock / John Lucke	NJ ("Pensauken problem")
13. 1950	O.D. Von Engeln	Central Finger Lakes, NY
14. 1951	John Hack / Paul MacClintock	Chesapeake, MD (soils/stratigraphy)
15. 1952	Dick Goldthwait	Central OH (tills)
16. 1953	Lou Currier / Joe Hartshorn	Ayer quad, MA (outwash sequences)
17. 1954	Charlie Denny / Walter Lyford	Wellsboro-Elmira-Towanda, PA-NY

18.	1955	Paul MacClintock	Champlain lake and sea, NY
19.	1956	Nelson Gadd	St. Lawrence Lowland, QUE
20.	1957	Paul MacClintock / John Harris	St. Lawrence Seaway, NY
21.	1958	John Hack / John Goodlett	Appalachians, Shenandoah, VA
22.	1959	Alexis Dreimanis / Bob Packer	Lake Erie, ONT (till bluffs)
23.	1960	Ernie Muller	Cattaraugus Co., western NY
24.	1961	Art Bloom	SW Maine (marine clay; ice margins)
25.	1962	Cliff Kaye / Phil Schafer	Rhode Island (Charleston Moraine etc.)
26.	1963	Hulbert Lee	Lower St. Lawrence Lowland, QUE
27.	1964	Cliff Kaye	Martha's Vineyard, MA
28.	1965	Joe Upson	Northern Long Island, NY
29.	1966	Nick Coch / Bob Oaks	Southeast VA (scarps; stratigraphy)
30.	1967	Hal Borns	Eastern ME (moraines; glaciomarine)
31.	1968	Carl Koteff / Bob Oldale / Joe Hartshorn	Eastern Cape Cod, MA
32.	1969	Nelson Gadd / Barrie McDonald	Sherbrooke area, QUE
33.	1970	Dick Goldthwait / George Bailey	Mt. Washington area, NH
34.	1971	Gordon Connally	Upper Hudson Valley, Albany, NY
35.	1972	Art Bloom / Jock McAndrews	Central Finger Lakes, NY
36.	1973	Don Coates / Cuchlaine King	Susquehanna-Oswego Valleys, NY-PA
37.	1974	Bill Dean / Peter Duckworth	Oak Ridges-Crawford Lake, ONT
38.	1975	George Crowl / Gordon Connally / Bill Sevon / Les Sirkin	Lower Delaware Valley, PA
39.	1976	Bob Jordan / John Talley	Coastal Plain, DE

40.	1977	Bob Newton	Ossipee quad, NH
41.	1978	Denis Marchand / Ed Ciolkosz / Milena Bucek / George Crowl	Central Susquehanna Valley, NY
42.	1979	Jesse Craft	NE Adirondack Mtns., NY
43.	1980	Bob LaFleur / Parker Calkin	Upper Cattaraugus, Hamburg, NY
44.	1981	Carl Koteff / Byron Stone	Nashua Valley, MA
45.	1982	Pierre LaSalle / Peter David / Michel Bouchard	Drummondville, QUE
46.	1983	Woody Thompson / Geoff Smith	Augusta-Waldoboro area, ME
47.	1984	Peter Clark / J.S. Street	St. Lawrence Lowland, NY
48.	1985	Ed Evenson / Jim Cotter / Dave Harper / Carl Koteff / Jack Ridge / Scott Stanford / Ron Witte	Great Valley, NJ-PA
49.	1986	Tom Lowell / Steve Kite	Northernmost ME
50.	1987	Carl Koteff / Janet Stone / Fred Larsen / Joe Hartshorn	Connecticut Valley-Lake Hitchcock, CT-MA
51.	1988	Ernie Muller / Duane Braun / Bill Brennan/ Dick Young	Genesee Valley, NY
52.	1989	Pierre LaSalle / Andree Blais / Denis Demers / Michel Lamothe / Bill Shilts	Mid St. Lawrence Lowland, QUE
53.	1990	Ralph Stea / Bob Mott	Halifax region, NS
54.	1991	Jack Ridge	Western Mohawk Valley, NY
55.	1992	Bob Dineen / Eric Hanson / Bob LaFleur / Dave Desimone	Lower Mohawk Valley, NY
56.	1993	Carol Hildreth / Richard Moore	Contoocook-Souhegan-Piscataquog Valleys, NH
57.	1994	Duane Braun / Ed Ciolkosz / Jon Inners / Jack Epstein	Eastern PA
58.	1995	Woody Thompson / Tom Davis / John Gosse / Bob Johnston / Bob Newton	Portland-Sebago Lake-Ossipee Valley region, ME