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Due to time and distance problems of the authors, a road log for this field trip is being issued separately. For similar reasons and because exposures of glacial deposits in this area are so ephemeral, specific field trip stops are not identified in this article. Detailed descriptions and interpretations are included, however, for the most likely stops. We intend to visit at least one glacial lake spillway, an exposure of collapsed lake sediments overlain by noncollapsed stream-terrace deposits, and exposures of deltaic deposits that pertain to the problems of lake damming, lacustrine and bog sedimentation, and postglacial uplift. Other appropriate stops can be made as time permits.

Introduction

This trip is concerned chiefly with deglacial meltwater features resulting from the last glaciation, but a few comments on earlier glacial history are in order. Although there were several Pleistocene glaciations that probably covered the region, solid evidence for just two major ice advances in New Hampshire has been demonstrated by detailed field work in recent years (Koteff, 1970; Newton, 1978; Koteff and Pessl, in press). Newton (1978) discusses the possibility of a third glaciation, although the evidence is very fragmentary. At that, the older of the two recognized glaciations is known only from scattered exposures of a compact, relatively silty till that is oxidized to a depth of as much as 25 feet (7.6 m). In contrast to the older (lower) till, the younger (upper) till is very widespread in distribution, is more sandy, and has been only slightly affected by oxidation. Therefore, the upper till is considered to be late Wisconsinan in age. Dating of the lower till is more problematical, and at present it is considered to be either early Wisconsinan or older, possibly Illinoian in age.

Deglaciation of the Merrimack Valley by the late Wisconsinan Laurentide ice sheet was characterized almost entirely by systematic stagnation-zone retreat (Koteff and Pessl, 1981), as was the case for all of southern and most of central New England. In this concept, the retreating ice sheet is viewed as having had a fringe of stagnant ice, perhaps at most a few kilometers wide, that acted as a buffer between the live ice and the meltwater deposits, or morphosequences, which were deposited partly within but mostly beyond the stagnant zone of ice. The numerous morphosequences in the Merrimack Valley are chiefly ice-marginal deltas and related fluvial and lake-bottom meltwater deposits that indicate ice retreat took place, for the most part, in a glaciolacustrine environment. The ice-marginal deltas and related features mark many former and successively younger positions of the stagnant ice edge from Tyngsboro, Massachusetts, to Concord, New Hampshire, and beyond. Only

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one interruption in the more or less steady ice retreat occurred near Manchester, New Hampshire, where evidence of a minor readvance of no more than a few kilometers has been found (Stone and Koteff, 1979).

Glacial Lakes

Formation of glacial lakes

In previous years (Goldthwait, 1925), it was thought that the rising marine waters that followed the retreating ice margin during deglaciation inland from the New England coast had extended up the Merrimack Valley as far north as Manchester, New Hampshire. However, Goldthwait (1938) makes no mention of this premise in his later article on uncovering of New Hampshire by the last ice sheet, and White (1938), in a brief discussion of glacial lakes in the Merrimack Valley, indicates that the former water body located northward from Nashua, New Hampshire, was lacustrine. So apparently by this time, the idea that the late-glacial sea had occupied this part of the Merrimack Valley had been discarded. Detailed mapping in recent years (Koteff, 1970, 1976) shows that three separate and successively formed glacial lakes existed in this region of the Merrimack Valley (Figure 1): Lake Tyngsboro, which occupied a small area across the Massachusetts-New Hampshire border; Lake Merrimack, which extended from Nashua to Manchester; and Lake Hooksett, which extended from about 3 mi (4.8 km)south of the village of Hooksett to north of Concord. The history of these lakes is somewhat unusual in that each of the last two lakes formed at a slightly higher level than that of the preceeding one (Figure 2). In north-draining valleys in New England, successive glacial lakes formed at successively lower levels as the retreating and damming ice uncovered successively lower spillways or outlets. In the south-draining Merrimack Valley, however, a different situation existed. Lake Tyngsboro formed behind meltwater deposits, probably deltaic, that dammed the valley near the village of Tyngsboro, Massachusetts, when the ice front stood there (Figure 1). This lake emptied via a bedrock-floored spillway at a present day altitude of about 140 ft (43 m). When the ice margin had retreated about 2.5 mi (4 km) north of the Massachusetts-New Hampshire border, deposition of the last of a series of Lake Tyngsboro deltas completely filled a narrow part of the valley there. These deposits, which have an altitude of at least 185 ft (56 m), blocked meltwater flow and served as a dam for the newly formed glacial Lake Merrimack. At the same time, an adjacent north-south gap or spillway floored in bedrock was uncovered (Figure 1). Although this spillway, at an altitude of 170 ft (52 m) was at or possibly 3 ft (1 m) lower than the inferred level of Lake Tyngsboro at this location, its narrowness caused hydraulic raising of the Lake Merrimack level to an altitude of about 178 ft (54 m). This allowed meltwater of the newly formed and expanding Lake Merrimack an escape route around the damming delta deposits of the final stage of Lake Tyngsboro, rather than overtopping and eroding them. Thus, if it were not for the propitiously positioned spillway for Lake Merrimack, meltwater flow would have prevented Lake Tyngsboro deposits from completely blocking the valley, and Lake Merrimack would have been confluent with Lake Tyngsboro, rather than being at a higher level.

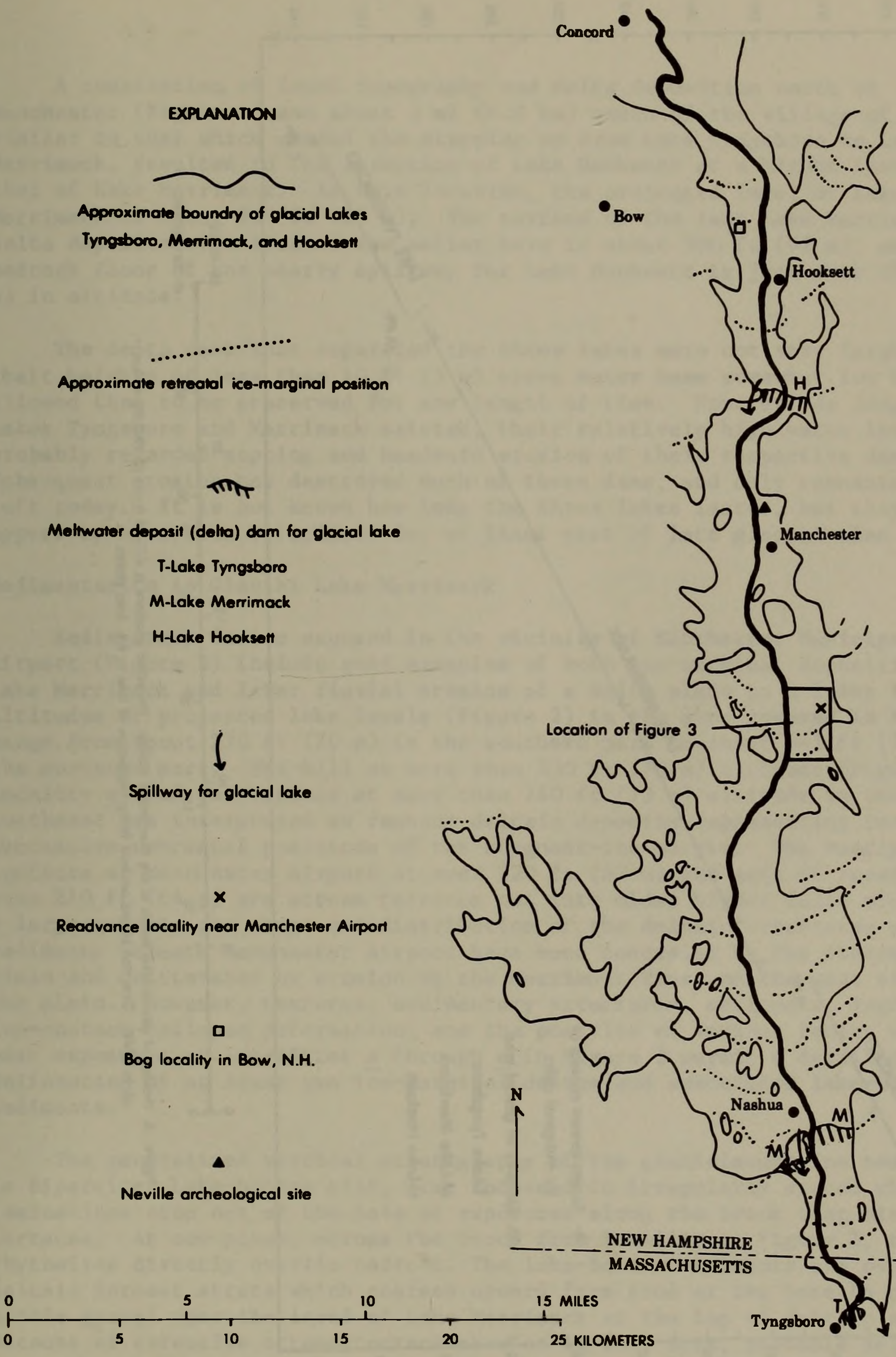


Figure 1. Outline of glacial lakes in part of the Merrimack Valley, Massachusetts—New Hampshire.

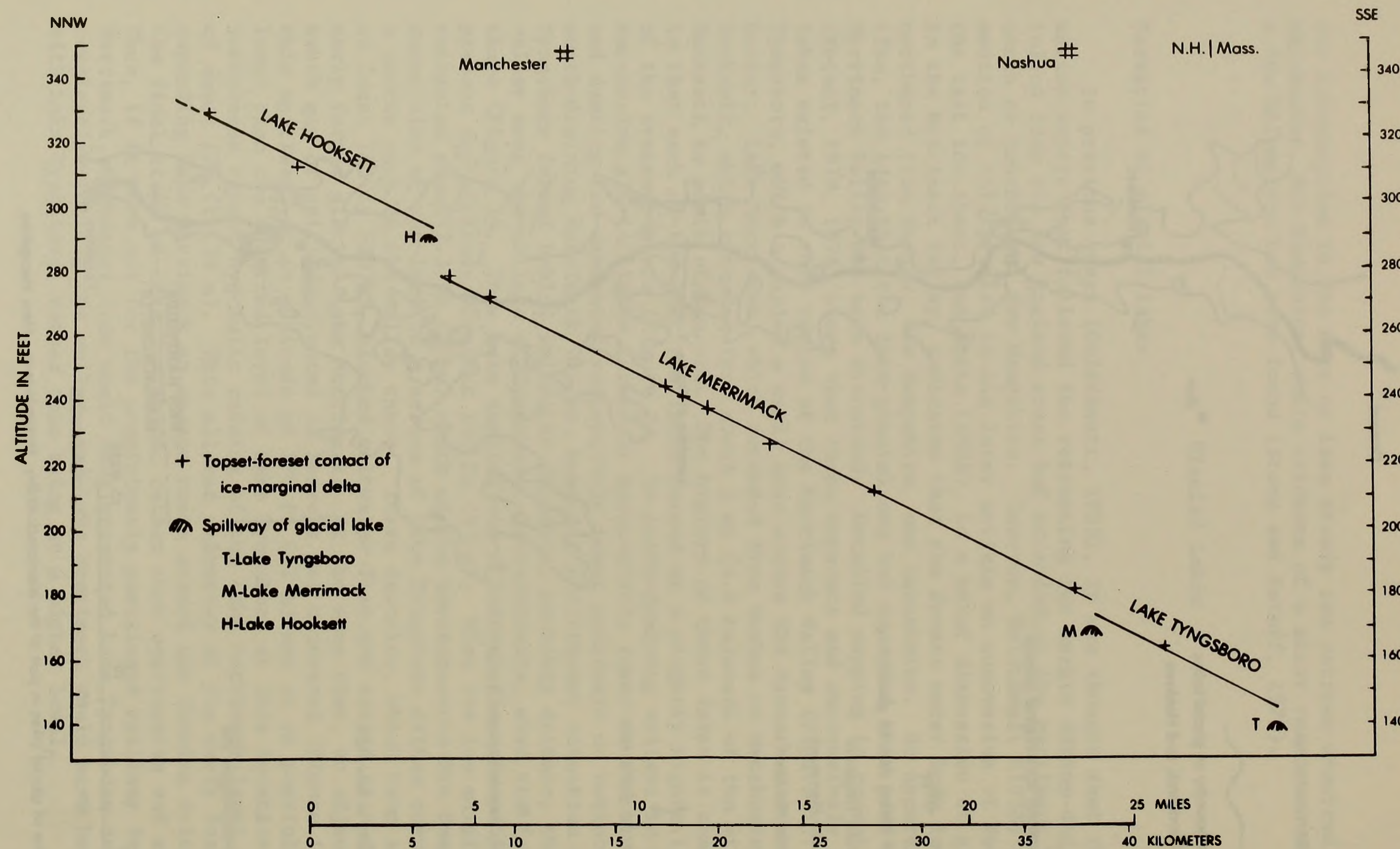


Figure 2. Profiles of glacial-lake levels in the Merrimack Valley. Projected gradient is 4.7-4.9 ft/mi (0.89-0.93 m/km) up to NNW. Direction based on best fit by inspection of line to the data points.

A combination of local topography and delta deposition north of Manchester (Figure 1) and about 3 mi (4.8 km) south of the village of Hooksett similar to that which caused the stepping up from Lake Tyngsboro to Lake Merrimack, resulted in the formation of Lake Hooksett at a higher level than that of Lake Merrimack. At this location, the projected level of Lake Merrimack is about 280 ft (85 m). The surface of the last Lake Merrimack delta deposits that blocked the valley here is about 300 ft (91 m), and the bedrock floor of the nearby spillway for Lake Hooksett is just over 290 ft (88 m) in altitude.

The delta dams that separated the three lakes were not very large, and their heights of less than 10 ft (3 m) above water seem somewhat low to have allowed them to be preserved for any length of time. However, as long as Lakes Tyngsboro and Merrimack existed, their relatively high water levels probably retarded sapping and headward erosion of their respective dams. Subsequent erosion has destroyed much of these dams, and only remnants are left today. It is not known how long the three lakes lasted, but they all appear to have been coincident for at least part of late glacial time.

Sedimentation in Glacial Lake Merrimack

Sedimentary facies exposed in the vicinity of Manchester Municipal Airport (Figure 3) include good examples of both ice-marginal deposition in Lake Merrimack and later fluvial erosion of a delta plain to a lower level. Altitudes of projected lake levels (Figure 2) in the area covered in Figure 3 range from about 230 ft (70 m) in the southern part to about 245 ft (75 m) in the northern part. The hill at more than 230 ft (70 m) altitude (Figure 3, locality a) and the terrace at more than 240 ft (73 m) altitude to the southeast are interpreted as remnant deltaic deposits representing two successive retreatal positions of the stagnant-ice margin. The nearly flat surfaces at Manchester Airport at over 220 ft (67 m) and east of locality a at over 210 ft (64 m) are stream terraces cut into older higher delta plains. To a large extent, the slope and distribution of the deltaic structures and sediments beneath Manchester Airport have been concealed by the stream-terrace plain and obliterated by erosion by the Merrimack River on the west side of the plain. However, textures, sedimentary structures, attitudes, degree of ice-contact collapse deformation, and the position of various facies seen in past exposures at localities a through e in Figure 3 permit a detailed delineation of at least two ice-marginal deltas and associated lake-bottom sediments.

The generalized vertical stratigraphy of the glaciolacustrine sediments is bipartite. Lake-bottom silt, clay and sand in irregularly spaced rhythmic laminations crop out at the base of exposures along the brook that bisects the terraces. At one place, across the brook from locality e, Figure 3, these rhythmites directly overlie bedrock. The lake-bottom sediments are overlain by deltaic foreset strata which coarsen upward from sand at the base to pebble-cobble gravel near the level of Lake Merrimack at the top of deltaic sections. Because of extensive stream-terrace erosion in this area, probable delta-topset beds of coarse gravel are preserved only locally (Figure 3, locality a). Shallow excavations in the surface of the plain around Manchester Airport exposed coarse pebbly sand and pebble gravel of the stream-terrace deposits that disconformably overlie the eroded deltaic beds. The textures of the terrace beds are similar to textures of well exposed, younger, lower, and

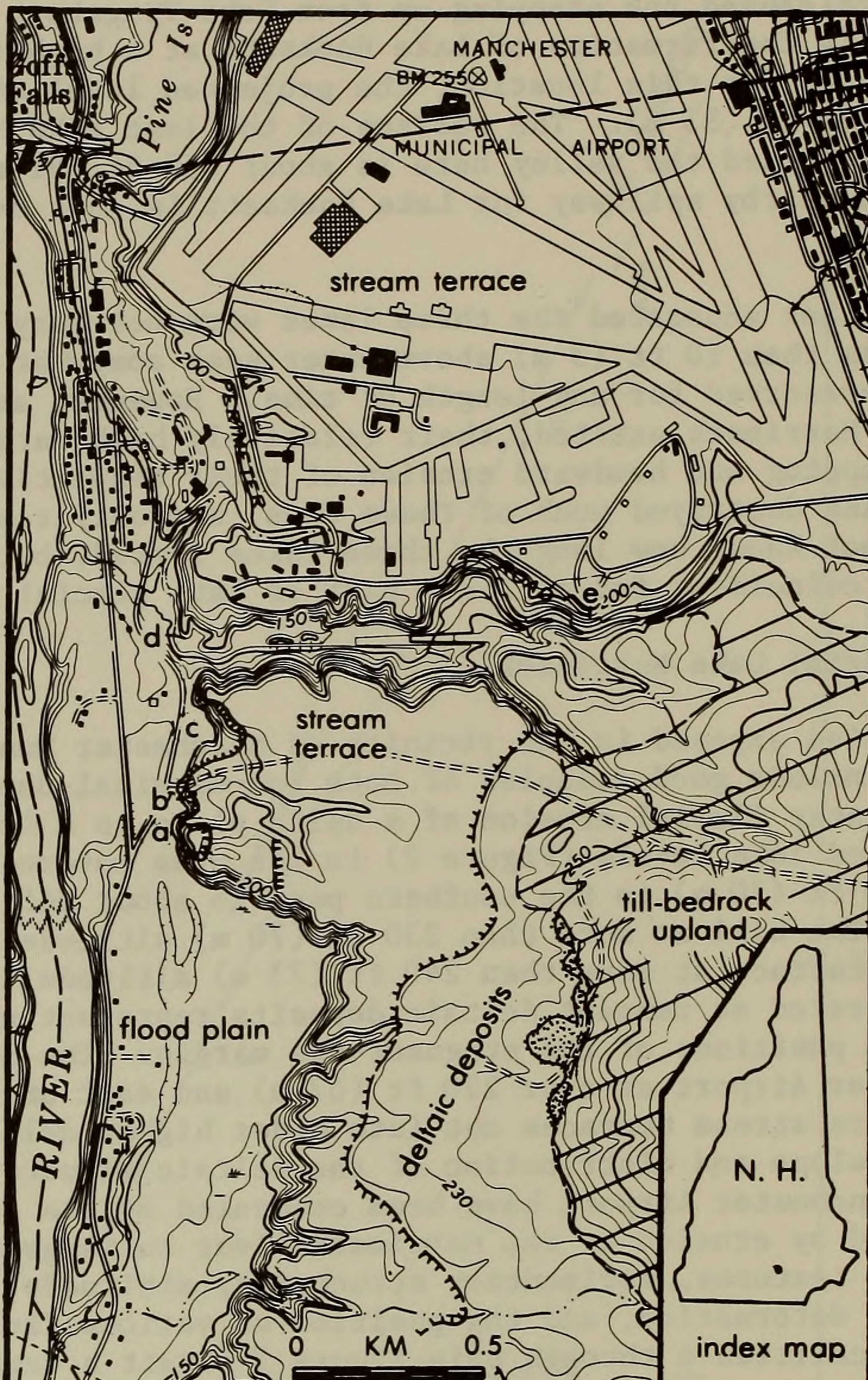


Figure 3. Morphology and stratigraphic localities of stream-terrace and glaciolacustrine deposits in the Manchester Municipal Airport area. Stippled pattern shows area of eolian sand deposits. Lettered localities referred to in text. Dashed ticked line indicates erosional scarp, ticks on downslope side. Base map from U.S. Geological Survey Manchester South 7¼′ quadrangle, 1968 edition. Continuous hachured line shows extent of pit excavations, 1970.

better defined stream-terrace deposits elsewhere in the valley. The terrace deposits are notably finer grained than gravel topset beds of ice-marginal deltas in the Merrimack Valley and of ice-marginal deltas in general (Stone, 1984), but are coarser than most delta foreset and lake-bottom sediments.

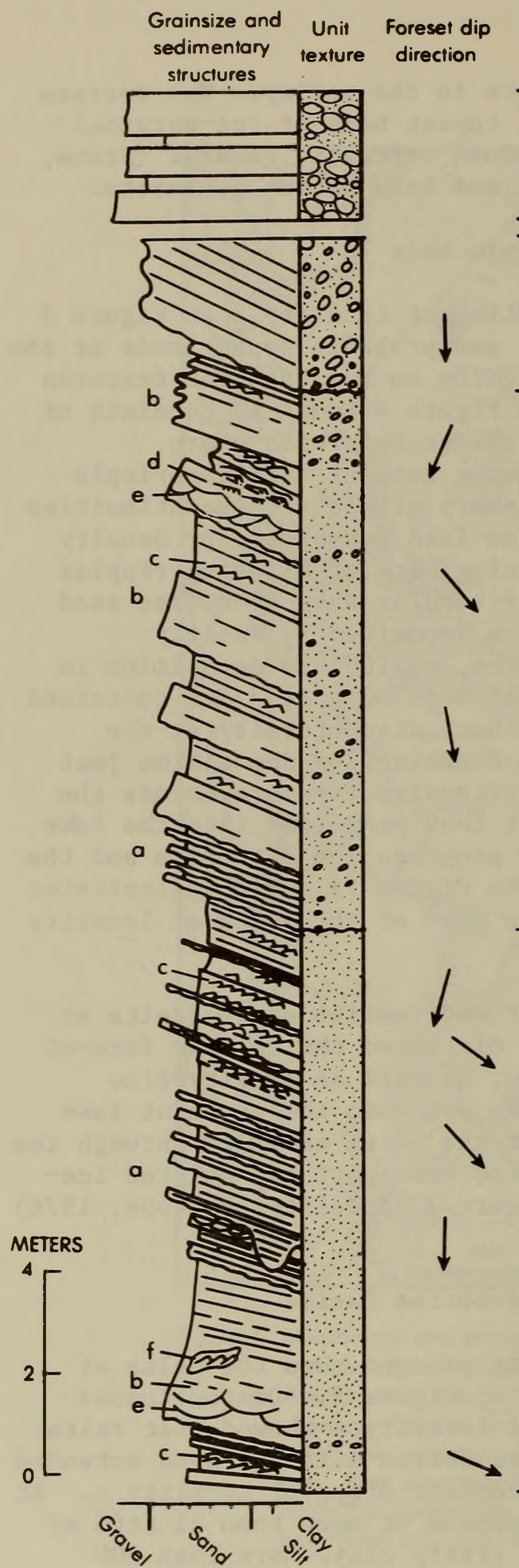
Vertical measured section in ice-marginal deltaic beds

Figure 4 is a composite stratigraphic section at locality a in Figure 3 showing coarsening-upward deltaic foreset beds and probable topset beds at the top. The foreset beds are divided into three units on the basis of textures and sedimentary structures. The lowest unit in Figure 4 (unit 1) consists of medium to coarse sand in laterally continuous planar beds with sharp, erosional lower boundaries, and in beds containing sets of climbing-ripple cross laminations. These structures, and the sharp erosional discontinuities between sets of foreset strata, suggest traction-load deposition by density underflows down slopes of 12-24°. Beds containing sets of climbing ripples and trough cross beds intercalated with locally tabular sets of coarse sand beds in unit 2 indicate that these foresets were deposited by density underflows, alternating with periods of avalanche, grain-flow deposition in foreset beds dipping 18-27°. The coarse foreset beds of unit 3 are contained in a tabular set of beds that terminates at a sharp disconformity at the base. These beds are interpreted as products of grain-flow deposition just beneath the surface of the lake. A laterally extensive exposure across the strike of these coarse foreset beds showed that they prograded into the lake along an azimuth of 178°. This upper foreset progradation direction and the range of other foreset dip azimuths are shown in Figure 5A, which illustrates the initial stages of progradation of the upper part of the delta at locality a, Figure 3.

A previously excavated zone of ice-contact deformation of the delta at locality a, Figure 3 showed normal faults that displaced the deltaic foreset strata down to the northwest. These structures, as well as gravity-flow deposits that flowed from the ice-contact slopes out onto the adjacent lake bottom indicate that the supporting ice wall of the delta extended through the pit at locality a. The direction of gravity-flow transport and related ice-contact faulting and gravity-flow deposits (Figure 3, locality b; Stone, 1976) are shown in Figure 5B.

Younger ice-marginal deltas and associated lake-bottom facies

Sandy foreset beds of an ice-marginal delta younger than the delta at locality a have been exposed in a pit locality c, Figure 3. Discontinuous exposures from the younger delta to the area of locality a showed that delta bottomset sand beds and distally equivalent lake-bottom silt and sand extended from the younger delta to the foot of the ice-contact slope of locality a. At locality a these younger bottomset beds are composed of more than 15 ft(5 m) of rhythmically laminated silt, fine sand, and gritty clay. More than 200 rhythmite couplets,averaging one inch (2.5 cm) in thickness, were exposed. Coarse layers of each couplet contain multiple laminations and microlaminations, local scour lenses, and ripple cross-laminations. The ripples indicate southeast paleocurrents, and the sediments presumably were derived from an active delta or lacustrine fan source to the northwest. The fine layers of the couplets are composed of gritty clay, are 1-7 mm thick, and occur as a drape in some places above ripples. At the base of the section,



Description of Units

Unit 4. Fluvial beds

Interbedded pebble-cobble gravel, pebbly coarse sand, and medium to coarse sand in horizontal beds.

Unit 3. Upper delta-foreset beds.

Pebble-cobble gravel and pebbly coarse sand in dipping tabular beds, shown schematically.

Unit 2. Middle delta-foreset beds.

Chiefly coarse sand and pebbly sand in dipping strata containing: (a) sets of parallel planar beds, (b) sets of planar laminae and very thin beds with indistinct bedding, (c) sets of B-type and (d) A-type climbing-ripple cross laminations, and (e) trough cross-beds.

Unit 1. Lower delta-foreset beds.

Medium to coarse sand in dipping strata containing: (a) sets of parallel planar beds, (c) sets of B-type climbing-ripple cross laminations, (e) trough cross-beds, (b) sets of planar laminae with indistinct bedding, and (f) intraclast of rippled sand.

Figure 4. Measured section of ice-marginal deltaic beds exposed at locality a, Figure 3. Figure modified from method of Selley (1970) to show delta foreset dips, bedding within unit sets of foreset strata, and unit textures.

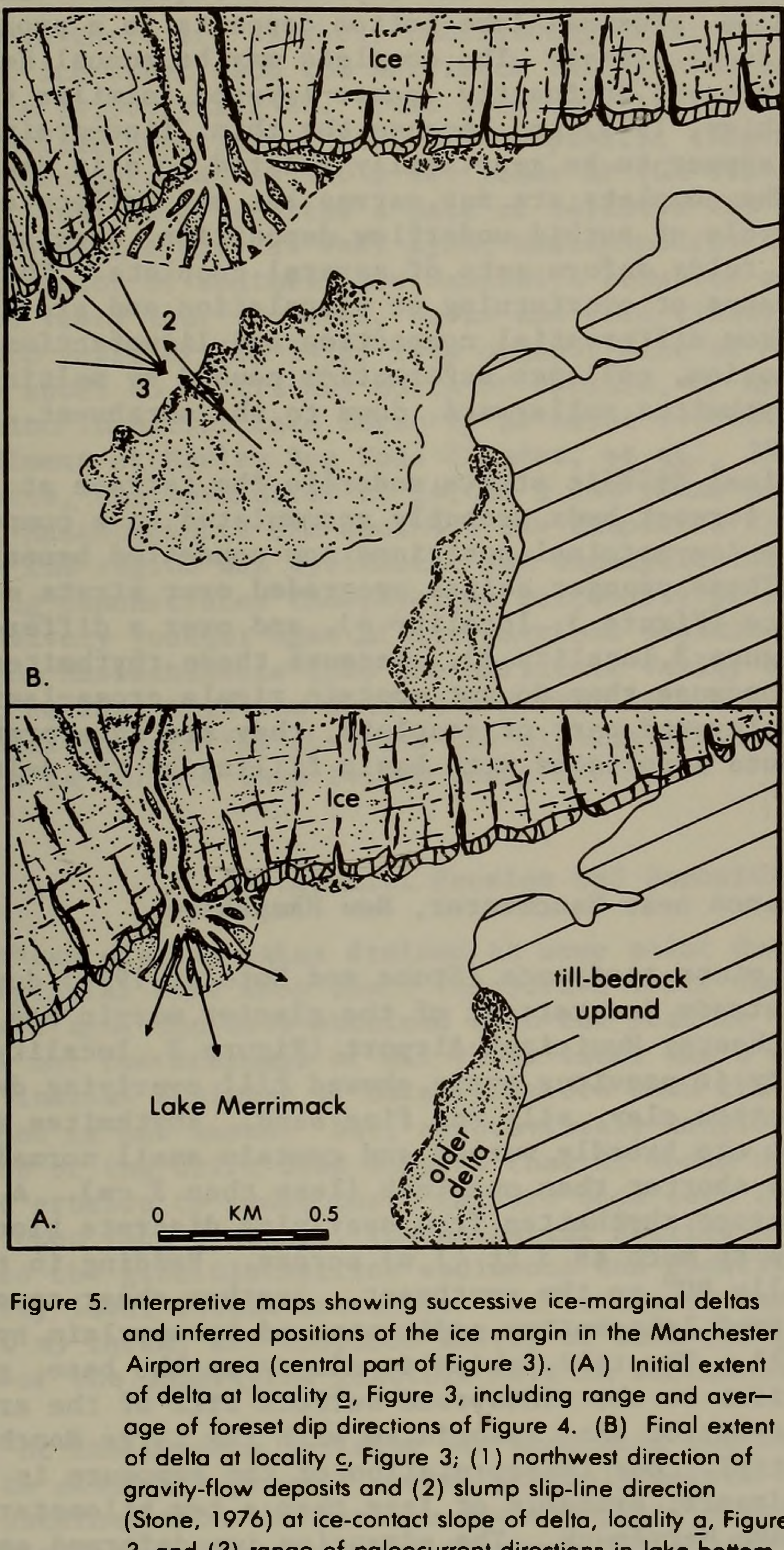


Figure 5. Interpretive maps showing successive ice-marginal deltas and inferred positions of the ice margin in the Manchester Airport area (central part of Figure 3). (A) Initial extent of delta at locality a, Figure 3, including range and average of foreset dip directions of Figure 4. (B) Final extent of delta at locality c, Figure 3; (1) northwest direction of gravity-flow deposits and (2) slump slip-line direction (Stone, 1976) at ice-contact slope of delta, locality a, Figure 3, and (3) range of paleocurrent directions in lake-bottom sediments, locality a, Figure 3.

gravity-flow deposits, composed of poorly sorted sand and containing silt-lamina intraclasts are intercalated within the rhythmite sequence. These rhythmites are interpreted as proximal lake-bottom deposits in a prodelta basin of probably limited local extent. The couplets may be annual deposits; however, because the couplets are dominantly coarse-layered (similar to rhythmite facies III of Ashley, 1975) and because the thin, fine-grained layers are silty clay and appear to be genetically related to silt draped laminae, we suggest that the couplets are not varves but may reflect a regular, perhaps diurnal cycle of turbid underflow deposition. Spectacular intraformational isoclinal folds deform sets of several couplets. These folds do not show a systematic sense of overturning or translation and are interpreted as resulting from differential compaction and liquefaction. Following this load deformation, collapse deformation caused by melting ice faulted and rotated the rhythmites valleyward, down to the northwest.

Even younger ice-marginal deltaic strata underlie the terrace at Manchester Airport. These foreset beds probably accumulated in a composite delta built from successive ice-marginal positions now concealed beneath the stream-terrace deposits. These younger deltas prograded over strata earlier deformed by an ice readvance (Figure 3, locality _e_), and over a different sequence of rhythmites (Figure 3,locality _d_). Because these rhythmites are not overlain by till, and because they do not contain ripple cross-laminations or relatively thick coarse-layered part of couplets, they are interpreted as younger lake-bottom sediments associated with basin filling by the deltas at the airport.

Ice readvance near Manchester, New Hampshire

Evidence for only one minor readvance (Stone and Koteff, 1979) during systematic and relatively steady ice retreat of the glacier margin has been found in the area near Manchester Municipal Airport (Figure 3, locality e). An exposure at this locality in previous years showed till overlying deformed rhythmically bedded lake-bottom clay, silt and fine sand. Rhythmites in the lower 3 ft (1 m) of section are broadly warped and contain small normal and thrust faults having traces shorter than one inch (less than 2 cm). A thin shear zone separates the warped rhythmites from overlying discrete blocks of rotated rhythmites that are as much as 3 ft (1 m) across. Bedding in the rotated blocks dips generally 80^{o} to the southeast. Another shear zone bounds the upper part of the deformed lake-bottom sediments and is overlain by as much as 5 ft (1.5 m) of till. The till, which is clayey at the base, grades upward into sandy till similar to the widespread surface till of the area. Undisturbed gravel and lake-bottom sand associated with the large Manchester Airport delta overlie the till. Our interpretation of the exposure is that it represents a local ice readvance, probably of less than a few kilometers, over lake-bottom sediments of Lake Merrimack. The advancing ice deformed and detached blocks of these sediments and overrode them, depositing till. The source of the clayey till was the local lake-bottom sediments, and the sandy portion represents ice transport from a greater distance. During subsequent ice retreat, deposits of the Manchester Airport delta covered the readvance locality, and systematic and steady ice retreat once again characterized deglaciation of the valley.

Age of Deglaciation

Although there are slightly conflicting data, the retreat of the last Wisconsinan ice sheet from the Merrimack Valley area is thought to have taken place before 14,000 years BP. Peat obtained from a kettle that formed in Lake Hooksett deltaic deposits at Bow, New Hampshire, (Figure 1; Caldwell et al., 1978) has been dated at 13,050 ± 325 years BP (GX-4554), and Kaye and Barghoorn (1964) have reported a date of 14,250 ± 250 years BP (W-735) for barnacles in marine clay at West Lynn, Massachusetts. These dates suggest that deglaciation of southern New Hampshire occurred near or slightly later than 14,000 years ago. However, Davis and Ford (1982) have reported a date of 13,870 ± 560 years BP (GX-5429) from organic silt at Mirror Lake, New Hampshire, about 45 mi (72 km) north of Concord, N. H. They feel that sedimentation in Mirror Lake began by at least 14,000 years ago Also, a limnic sediment at Hawley Bog Pond (Bender, et al., 1981) near Plainfield, Massachusetts, has been dated at 14,000 ± 130 years BP (WIS-1122). Detailed mapping of Massachusetts and southern New Hampshire clearly indicates that ice retreat in the region was systematic and uncomplicated by major readvances. This mapping demonstrates the physical continuity of meltwater deposits and the successively younger ages of ice-marginal positions of the retreating glacier from Massachusetts into the Merrimack Valley of New Hampshire. Therefore, we feel that the data from Mirror Lake and Hawley Bog Pond provide a more realistic date for at least the minimum age for deglaciation in the area, which is earlier than 14,000 years BP.

Postglacial Erosion and Deposition

All three glacial lakes drained at some point during late glacial and early postglacial time when their respective drift dams failed, probably after 13,000 years ago, the date obtained from the peat at Bow, New Hampshire. Whether or not the drainage of all three lakes was a domino effect set in motion by initial draining of Lake Tyngsboro when the deltaic dam at Tyngsboro was breached is not known. But, as previously mentioned, the small size and low heights of the drift dams suggest that it would have required very little subsequent erosion to cause their failure once each lake to the south drained. Upon lake drainage, the ancestral Merrimack River began eroding deeply into the glaciolacustrine sediments and constructing a series of successively lower stream terraces. The stream-terrace deposits, as much as 15 ft (4.6 m) thick, are composed mostly of sand and gravel, and are coarser than most of the underlying glaciolacustrine sediments.

Most of the terrace building was completed and the Merrimack River had cut down to about its present course probably before 8,000 years ago. Charcoal obtained from the Neville archeological site located on a stream-terrace (Dincauze, 1976) near the Amoskeag Bridge in Manchester, New Hampshire (Figure 1), has yielded a radiocarbon date of 7740 ± 280 years BP (GX-1746). This terrace is the lowest stream-terrace above the modern floodplain, and the Merrimack River there flows on bedrock over a rapids, referred to locally as the Amoskeag Falls. Dincauze (1976) concludes that the Neville site was occupied by people attracted by migrating fish that could be taken at the falls. Thus, the Merrimack River appears to have been established by that time in its present course, at least in this part of the valley and probably for most of its length.

During downcutting by the Merrimack River and the formation of stream terraces, there was intensive eolian activity in parts of the valley, particularly on the east side between Nashua and Manchester. Large sand dune deposits as much as 40 ft (12 m) thick were formed predominantly by winds from the northwest. The dunes, which are composed of fine to medium well-sorted sand, occur as transverse and longitudinal dunes as well as amorphous masses. The sand was derived from glaciolacustrine sediments and probably from the higher stream-terrace deposits also. No dunes occur on the very lowest stream terraces and the modern floodplain.

Postglacial Uplift

Features related to glacial Lakes Tyngsboro, Merrimack, and Hooksett have provided evidence for the isostatic response of the crust that resulted from unloading of the late Wisconsinan ice sheet. Altitudes of the topset-foreset contact of ice-marginal deltas deposited in the lakes and of their spillways show a north-northwest trending profile (Figure 2) of uplift for this area that has a gradient of 4.7 to 4.9 ft/mi (0.89 to 0.93 m/km) up to the north. The north-northwest trend of the profile is an approximate best fit done by inspection of the data points; none of the projected profiles of the three lakes is long enough, and there too few altitudes of topset-foreset contacts of deltas of Lakes Tyngsboro and Hooksett, to allow a more precise determination of uplift direction at this time. The altitude of topset-foreset contacts shown in Figure 2 are believed to represent the lake level to within 3 ft (1 m), but the spillway floors are only the minimum lake altitude. A depth of water that varies in height from 4 to 8 ft (1.2 to 2.4 m), depending on the particular spillway, can be projected from the topset-foreset altitudes to the thresholds. The profiles shown in Figure 2 are based only on altitudes of noncollapsed parts of ice-marginal deltas. Many points that fall below these projections have been obtained from apparent topset-foreset contacts but these have been found to represent either collapsed topset-foreset contacts or tops of delta foreset beds that have been modified by later erosion. In the eroded deltas, later stream-terrace fluvial deposits overlie eroded foreset beds, mimicking topset-foreset contacts. As previously mentioned, these stream-terrace beds generally are finer grained than delta topset beds in the Merrimack Valley. Because the timing of the modification by erosion is not well known, data points from the eroded deltas can not be connected with any confidence and are not shown.

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