Quaternary Geology of Connecticut

Illustrated by a fieldtrip in the central Connecticut Valley

Geological Society of Connecticut

Fieldtrip Guidebook No. 4

Cover photo: Ice-dammed pond deposits in Dark Hollow Brook valley, Glastonbury, CT

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INTRODUCTION

The purpose of this fieldtrip is to demonstrate some of the important geologic deposits and features formed in Connecticut during the Quaternary Period, which includes the Pleistocene (glacial) and Holocene (postglacial) Epochs. The Quaternary Period has been the time of development of many details of the landscape and of all the surficial deposits. These materials are of pervasive importance to the present occupants of the land. Three main types of surficial deposits are present in Connecticut: 1) glacial ice-laid deposits were laid down beneath ice as continental ice sheets advanced across Connecticut at least twice in the Middle and Late Pleistocene; 2) glacial meltwater deposits were deposited in glacial lakes and streams as the last (Late-Wisconsinan) ice sheet retreated northward across Connecticut; and 3) postglacial deposits formed by various processes after the recession of the last ice sheet, many of them during the time in which the land rebounded from glacio-isostatic depression produced by the great weight of the continental ice sheet.

Glacial and postglacial deposits and features across Connecticut have been mapped and are illustrated on two State geologic maps. The Quaternary Geologic map of Connecticut and Long Island Sound Basin (Stone and others, 2005) illustrates the geologic history and the distribution of depositional environments during the emplacement of glacial and postglacial surficial deposits and the landforms resulting from those events. A companion map, the Surficial Materials Map of Connecticut (Stone and others, 1992) emphasizes the surface and subsurface texture (grain-size distribution) of these materials. The features portrayed on the two maps are very closely related; each contributes to the interpretations of the other.

On this fieldtrip we will examine glacial features and landforms and their internal structure in the central Connecticut towns of Glastonbury, Rocky Hill, Portland, Middletown, and New Britain.

GLACIAL ICE-LAID DEPOSITS

During the last (Late-Wisconsinan) glaciation (28-23 ka), a sector of the Laurentide ice sheet of northeastern North America spread across the St. Lawrence River valley and the Green and White Mountains of Vermont and New Hampshire, covered all of Connecticut, and reached its maximum extent on Long Island, New York. Ice-movement directions are indicated by striations and grooves on bedrock, drumlin axes, and, inferentially, by the positions of ice margins during retreat. Ice movement across the State was dominantly from north-northwest to south-southeast (Fig. 1). The principal departure from that general trend was a prominent lobation in and adjacent to the Central Lowland. On the western side of that lobe, which probably became accentuated as the ice thinned during retreat, directions of movement were to the southwest or even to the west. Weaker lobate patterns occurred in the valleys of the Quinebaug, lower Connecticut, and Housatonic Rivers, and ice movement in westernmost Connecticut was influenced by the large lobe in the Hudson River valley. Less is known of the earlier (Illinoian) glaciation recorded by the presence of a lower till. Drumlins are composed dominantly of this lower till, and their axial directions are probably partly inherited from the earlier glaciation. Glacial meltwater deposits of this earlier glaciation are rare; they evidently were eroded or buried during the late Wisconsinan glaciation. Still earlier continental glaciations, recorded by deposits in the mid-continent and confirmed by oxygen-isotope studies in ocean sediments and Greenland ice cores (Imbrie and others, 1984; Mix, 1987; Paterson and Hammer, 1987), probably also affected Connecticut, but no direct evidence has yet been found within the State.

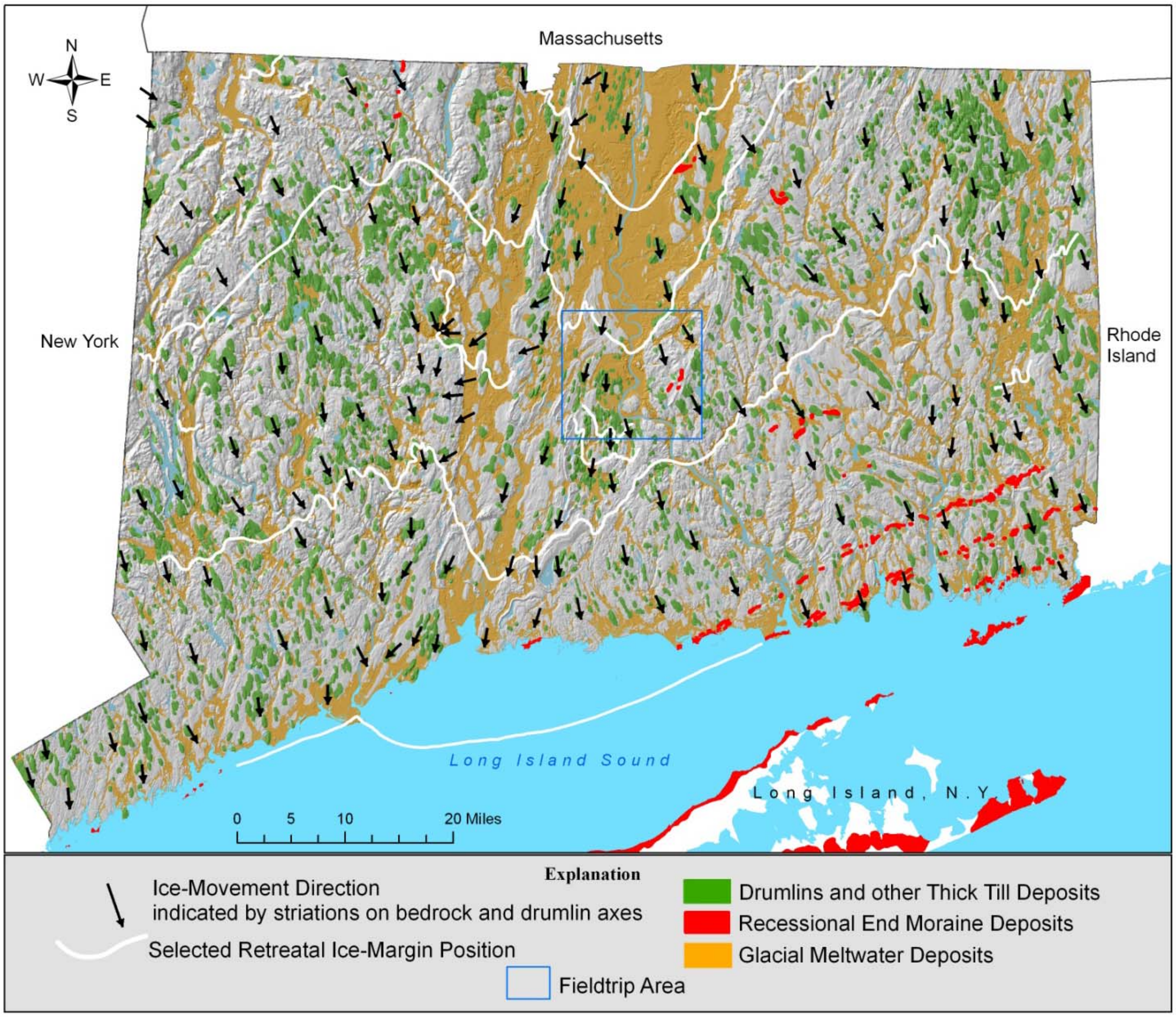


Figure 1. Ice-movement directions, distribution of drumlins and other thick till areas, recessional moraines, and glacial meltwater deposits in Connecticut (modified from Stone and others, 2005).

Glacial ice-laid deposits include till deposits and end moraine deposits (Figure 1). Glacial till was deposited beneath or melted out of glacial ice and is a nonsorted, stony mixture of sand, silt, clay, and boulders. The color, texture, and composition of till is related closely to the underlying and northerly adjacent bedrock units from which till was derived. Till deposits in the eastern and western highlands are light-gray sand to silt-sand, containing clasts of gneiss, schist, granitic rocks, minor quartzite, and local mafic rocks. Tills in the central lowland are dark-reddish-brown to yellowish-brown, silty to clayey silt-sand containing clasts of sandstone, basalt, diabase, and erratic clasts of gneiss, schist, and quartzite. In valleys underlain by marble in the western part of the State, tills are silt-sand with calcareous matrix containing clasts of marble, quartzite, schist, and gneiss.

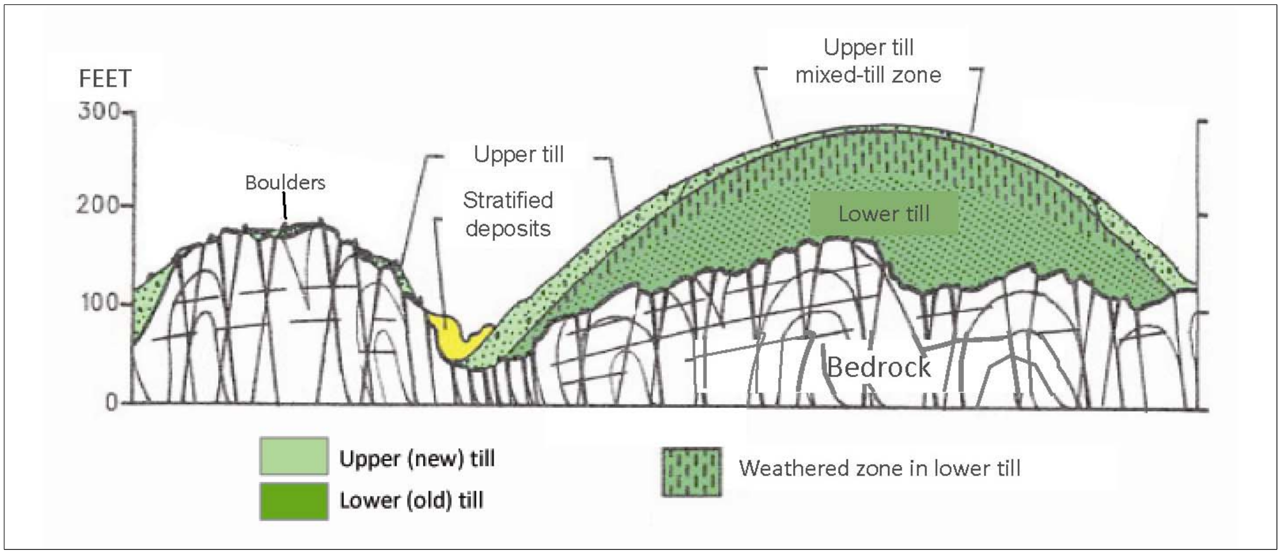


Figure 2. Idealized cross-section of a typical drumlin in southern New England showing the distribution of two distinct tills of different ages. Modified from Melvin and others, 1992.

Two glacial tills, distinctive in character and different in age, occur in superposition in Connecticut. The upper, younger till (also called surface till) was deposited by the late-Wisconsinan ice sheet and is a loose to moderately compact, generally sandy, commonly stony material consisting of both lodgement and ablation facies. Upper till is the most extensive till deposit and it is commonly exposed in shallow surface excavations, especially in areas where till thickness is less than 15 ft (light gray areas on Figure 1). The lower, older till (also called drumlin till) was deposited during the earlier Illinoian glaciation and is moderately to very compact, commonly finer grained and less stony than upper till. Lower till is less areally extensive than upper till, is primarily a subsurface unit, and generally is overlain by upper till (Fig. 2). Lower till constitutes the bulk of material in areas where till thickness is greater than 15 ft (green areas on Figure 1). Where both tills are exposed together, the base of upper till truncates the weathered surface of older till. An oxidized zone (lower part of soil profile developed during period of interglacial weathering) generally is present in the upper part of lower till section; this zone commonly shows closely spaced joints stained with iron and manganese oxides. The lower part of upper till commonly displays a zone of shearing and brecciation in which clasts of lower till are mixed and incorporated into upper till (Stone, 1989).

Generally the lower till is at the surface today only in floors of artificial excavations that are too small to map across the entire State. The lower till may be at the surface in the upper parts of some drumlins, but its occurrence is known only from local exposures and its extent cannot be predicted; most commonly, lower till in drumlins is mantled by thin, upper till. Numerous artificial exposures and subsurface well and test-boring data indicate that lower till constitutes the bulk of material within drumlins and other thick till areas.

End moraine deposits in Connecticut occur primarily in the southeastern part of the State, although there are scattered features present in other areas as well. These moraines are relatively low-relief, segmented boulder ridges (Fig 3.) that mark recessional positions of the retreating late-Wisconsinan ice sheet. The most extensive southeastern Connecticut moraines (Old Saybrook-Wolf Rocks, Hammonassett-Ledyard, and Madison-Oxoboxo) extend eastward into Rhode Island and westward offshore and are aligned with ridges of ice-marginal lacustrine fan deposits beneath Long Island Sound. The moraine belts are relatively linear, but show down-ice topographic deflection where they cross valleys. Accumulations of rock debris were concentrated in the shear zone where active ice rode up over thin stagnant ice at the margin. When the shear zone remained in one position for a significant time, concentrations of debris built up within and on top of stagnant ice (Goldsmith, 1982). This material was later deposited on the land surface by ablation processes. The linear trend of the moraine belts reflects the former position of the shear zone some relatively short distance behind the more ragged margin of stagnant ice. The segmented nature and local boulder-rich character of these moraines is probably due to the action of meltwater in the marginal zone. The moraine segments are most obvious in the upland areas between valleys. In valleys, morainic material may be buried by meltwater deposits; in most places, meltwater deposition dominated in the valley and the morainic position is represented by the ice-proximal head of a morphosequence. Locally, moraine segments, which are more lobate than in upland areas, stand at the proximal heads of meltwater deposits in the valley.

Figure 3. Bouldery section of the Hammonasset-Ledyard Moraine in Glacial Park off Whalehead Road, Ledyard, Connecticut. *Black Dog* for scale.

GLACIAL MELTWATER DEPOSITS

The shrinkage of the late Wisconsinan ice sheet and the retreat of its margin from south to north across Connecticut were accomplished as the ice melted faster than it was re-supplied by movement from the north. Meltwater picked up rock debris carried by the ice and deposited most of it shortly beyond the ice margin. The deposits are sorted, stratified layers of gravel, sand, silt, and clay; these sediments accumulated in streams and lakes, large and small, that were fed by the meltwater. Because meltwater largely flowed in valleys during deglaciation, meltwater deposits are concentrated in those valleys and in many places are more than 100 ft in thickness. Most of these thick meltwater sediments in Connecticut were deposited in or graded to large and small glacial lakes. Four types of glaciolacustrine depositional systems are present in Connecticut and are shown on Figure 4. Major glacial lakes (IL and SL depositional systems) dominate the deglacial history of the State. Deposits of these lakes, as well as those of the multitude of smaller glacial lakes (IP and SP depositional systems) and glaciofluvial systems (FP and FD depositional systems) record a detailed history of ice retreat across Connecticut.

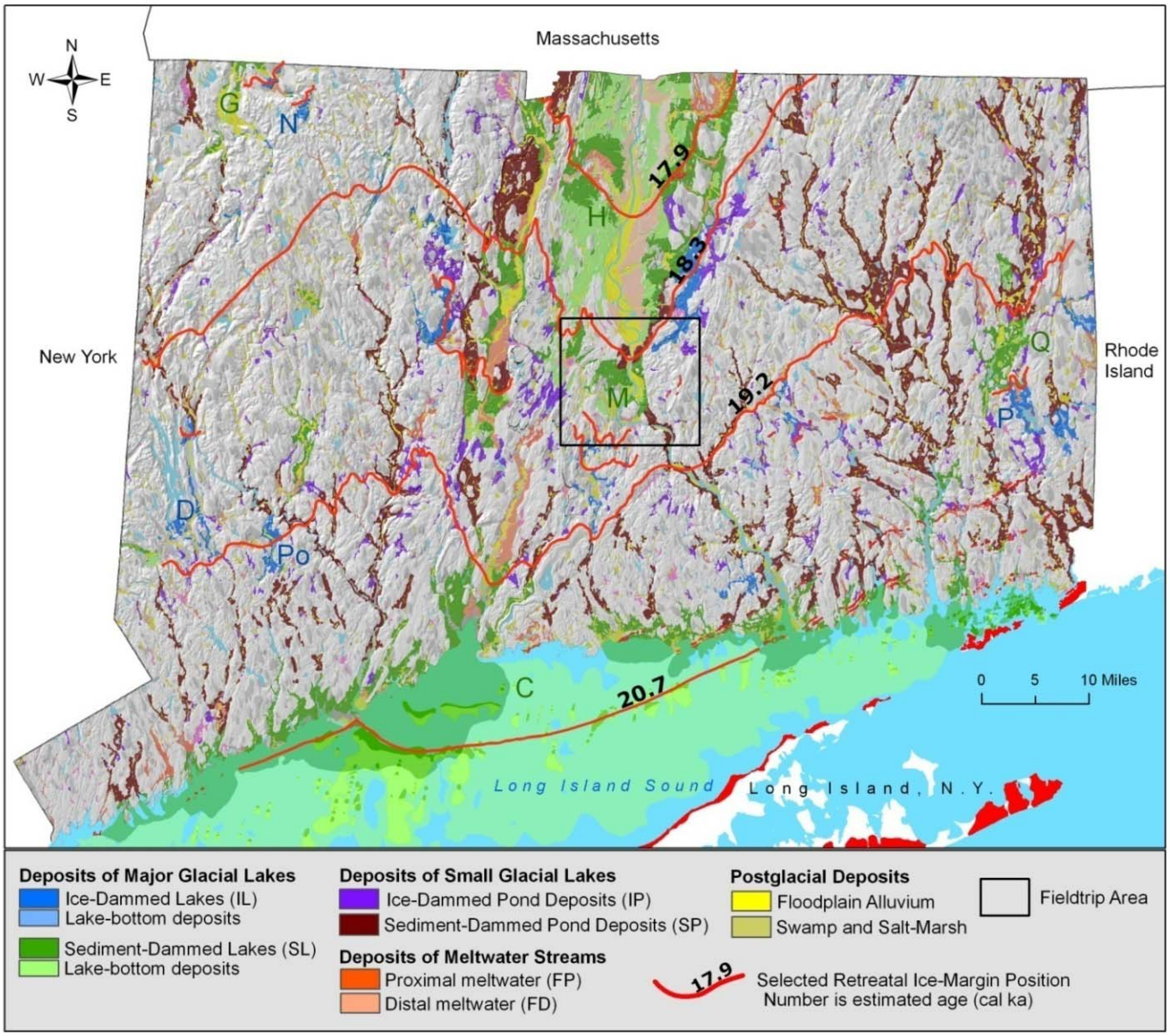


Figure 4. Glacial meltwater deposits characterized by depositional system and selected ice-margin retreatal positions in Connecticut. Major glacial lakes in the Connecticut Valley and Long Island Sound basin include: C, Glacial Lake Connecticut, M, Glacial Lake Middletown, and H, Glacial Lake Hitchcock; in the Western highlands, G, Glacial Lake Great Falls, N, Glacial Lake Norfolk, D, Glacial lake Danbury, Po, Glacial lake Pootatuck; in the Eastern highlands, Q, Glacial Lake Quinebaug, and P, Glacial Lake Pachaug.

Selected ice-margin retreatal positions are also shown on Figure 4. The approximate dates (given in calibrated years) associated with the four positions are estimated from ^{10}Be dates on boulders atop the Ledyard and Old Saybrook Moraines (Balco and others, 2009), the newly calibrated North American varve chronology (Ridge, 2013), and new varve sections from Glacial lake Middletown and Glacial Lake Hitchcock (Stone and Stone, 2012).

Sedimentary Facies and Morphosequences—Glacial meltwater deposits in Connecticut are composed of various sedimentary facies that are defined on the basis of lithic characteristics of texture and sedimentary structure and are related to specific environments of deposition along the path of meltwater flow: fluvial sediments were deposited in meltwater streams; deltaic sediments were deposited where meltwater streams entered glacial lakes; and lake-bottom sediments were deposited on the bottom of glacial lakes. Glacial sedimentary facies are combined either in facies assemblages or as single mappable bodies of sediment known as morphosequences (Koteff and Pessl, 1981). The types of morphosequences and the sedimentary facies included in them are described in Appendix 1. In general, a morphosequence is coarse grained at the glacier-proximal head and occurs in collapsed, ice-contact landforms; grain size decreases and landforms are less collapsed to noncollapsed in distal parts of the morphosequence. Morphosequences were deposited in close association with the ice margin; the surface altitude of each morphosequence was controlled by a specific base level, either a glacial lake plane or a valley knickpoint. Stratigraphic relationships between morphosequences in individual valleys provide ubiquitous evidence that these ice-marginal deposits are systematically younger from south to north (Fig. 5). Morphosequences are the basic mappable units of meltwater deposits at 1:24,000 scale, but they are too small and too numerous to map as individual units across the entire State, although ice-margin positions at the heads of most morphosequences have been mapped.

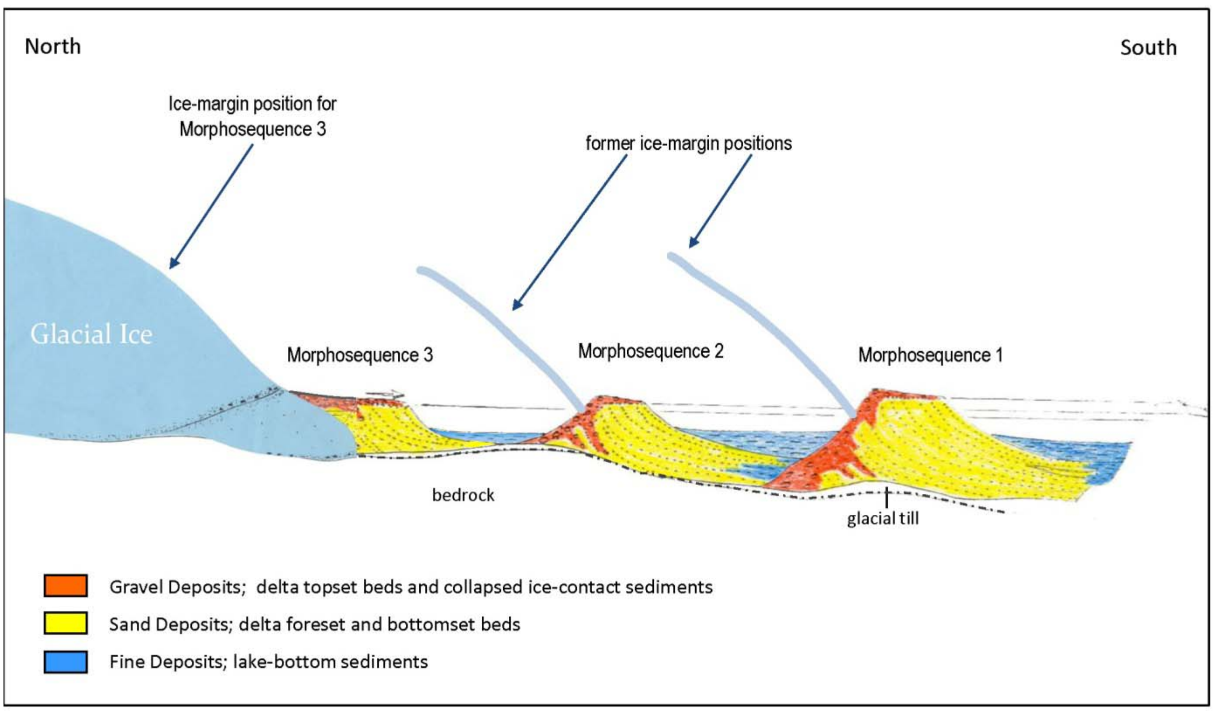


Figure 5. Sequential morphosequence deposition from south to north in a valley.

There are 204 correlated map units of meltwater deposits shown on the Quaternary geologic map of Connecticut. Each unit is a group of morphosequences deposited along the same or related paths of meltwater flow. Each unit was deposited either in a single glacial lake, a related series of lakes, or along meltwater streams in a valley where no ponding occurred. The position of groups of morphosequences (map units) in the landscape further indicates the systematic northward retreat of the ice margin. Where drainage divides were parallel or oblique to the trend of the ice margin, groups of high-level deltaic sediments were deposited when paths of meltwater escape were first held to higher positions against or through uplands, and then gradually lowered as lower paths were uncovered in valleys. On the basis of stratigraphic relationships between deposits, successive retreating ice-margin positions, and changes in glacial lake levels and in paths of meltwater flow, morphosequences are grouped into map units that are chronostratigraphic in character and that define a relative chronology of ice retreat across the State.

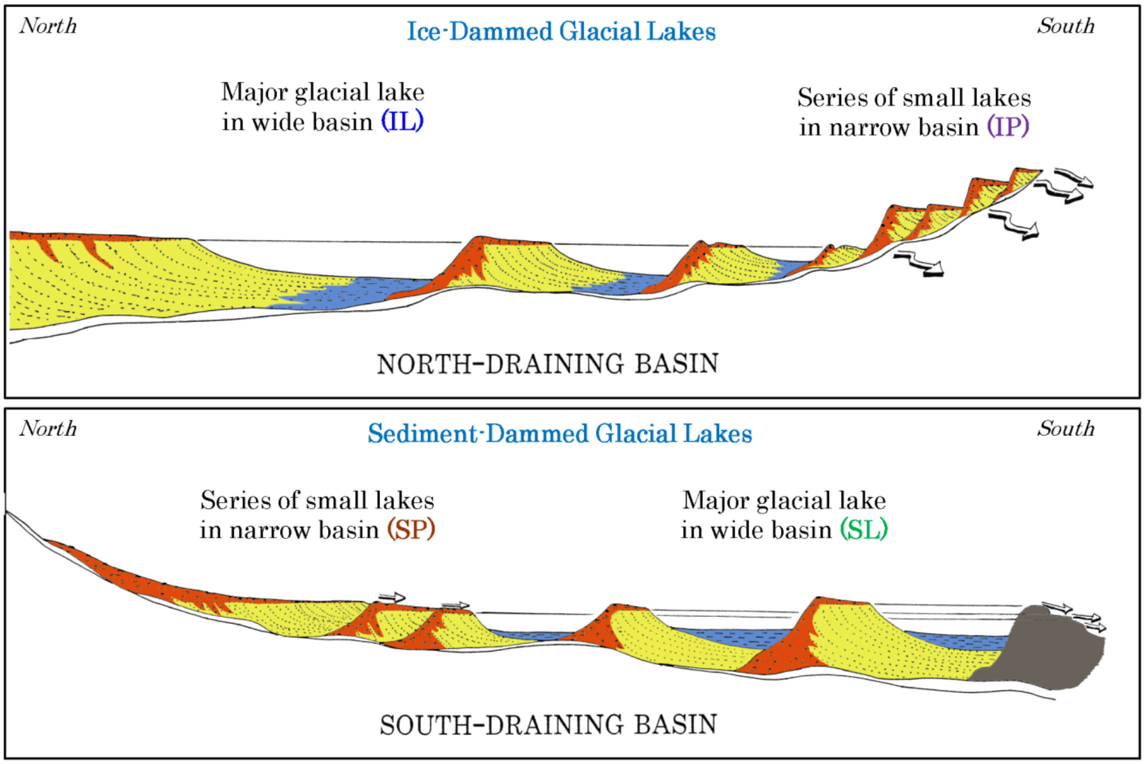


Figure 6. Glacial Lake depositional systems in Connecticut

Glaciolacustrine Systems-- Sediments of four glaciolacustrine systems (IL, SL, IP, and SP) were deposited in or graded to glacial lakes and ponds (Fig. 6). Sediments deposited in lakes include delta foreset and bottomset beds, lake-bottom sediments, and local lacustrine fan sediments. Sediments graded to glacial lakes include fluvial delta topset beds, delta-tributary fluvial sediments (fluvial sediment graded to deltas), and local ice-channel sediments. Deposits of glaciolacustrine systems are predominantly deltaic. Altitudes of topset-foreset contacts in deltas record the paleo-water-plane altitudes of the glacial lake into which they were built. Deltas of all glacial lakes in Connecticut indicate paleo-water-plane slopes of 4.7 ft/mi to the north-northwest. This slope is due to the glacio-isostatic tilt of the Earth's crust.

Large glacial lakes include ice-dammed lakes (IL) and sediment-dammed lakes (SL). The respective map units include all sediments graded to or deposited in single, relatively large, specifically named glacial lakes, some of which had several stages. These lakes existed in the wider valleys and large basins of the State. Deposits of major glacial lake systems are distinguished by several morphologic and stratigraphic characteristics: (1) deltas in each glacial lake (or lake stage) are at similar altitudes (when adjusted for glacio-isostatic tilt); (2) deltas have free fronts (that is, they prograded outward into the lake without being obstructed by earlier deposits and grade into flat-lying lake-bottom sediments); (3) lake-bottom deposits occur in front of the deltas; and (4) delta-tributary fluvial deposits occur in side valleys (valleys that were tributary to the glacial lake).

Small glacial lakes include ice-dammed ponds (IP) and sediment-dammed ponds (SP). The respective map units include all sediments graded to or deposited in sequentially ponded and chronologically related series of small lakes (ponds). These small lakes existed in the narrower valleys and small upland basins of the State. Deposits of small glacial lake systems are distinguished by several morphologic and stratigraphic characteristics: (1) deltas in each map unit are at divergent altitudes; (2) deltas commonly do not have free fronts, but rather are contiguous with each other; (3) lake-bottom sediments occur only beneath the deltas, not at the surface; and (4) fluvial deposits (only in depositional system SP) occur in steeper sections of the main valley and sometimes overlie deltaic deposits.

Glaciofluvial Systems-- Sediments of two glaciofluvial systems (FP and FD) were deposited in meltwater streams that were not tributary to any glacial lake. Meltwater streams deposited ice-marginal and near-ice-marginal glaciofluvial sediments in the steeper sections of some south-draining valleys and in front of moraines; these are deposits of proximal meltwater streams (FP). Sediments of distal meltwater streams (FD) were deposited in other valleys after glacial lakes in those valleys had drained. These deposits are generally thin and cap glacial lake sediments on meltwater terrace surfaces incised into the lake beds.

LARGE GLACIAL LAKES IN THE CENTRAL CONNECTICUT VALLEY

Two major glacial lakes dominated the landscape in the central Connecticut valley during ice retreat--glacial Lake Middletown and glacial Lake Hitchcock.

Glacial Lake Middletown first developed along the Connecticut River and in the Mattabesset River basin. The lake was impounded by a long mass of earlier deposits in the lower Connecticut River valley at and south of The Straits; the spillway, with an initial altitude of about 130 ft was over these deposits. Successive ice-marginal deltaic deposits were built into the lake as the ice retreated northward. When adjusted for the regionally established postglacial tilt of 4.74 ft/mi to N 21° W, delta topset-foreset contacts indicate that the lake slowly lowered due to erosion of its sediment dam. Glacial Lake Middletown occupied the Middletown basin in the lower Mattabesset valley and extended into the Berlin basin in the upper Mattabesset valley, as indicated by accordant delta levels, by basin geometry resulting in ice-margin positions that trend northwest-southeast, and by the extent of clays in the Berlin area. Deltas in Cromwell, Newington, and New Britain were built contemporaneously and record lake levels at the spillway of about 110 to 115 ft. Recently (2010) lake-bottom varved silts and clays were cored at two sites in the Newfield section of Middletown (see Stop 7). In the deepest core hole, 75 ft of varved clay contained 171 varves and indicates that Lake Middletown persisted for about 200 years during the time of delta building.

Just north of the Cromwell deltas, deltas of the Dividend Brook deposits were laid down in waters that were temporarily ponded to a higher level than glacial Lake Middletown and were controlled by the Dividend Brook spillway over Cromwell deltaic deposits; this spillway was not eroded lower than its present level of 129 ft because of the presence of glacial Lake Middletown at its mouth.

When the ice uncovered the lower part of the divide between the Hartford basin and the Middletown-Berlin-New Britain basin, where the New Britain spillway of glacial Lake Hitchcock would later exist, glacial Lake Middletown persisted at a level high enough to spread across the divide into the Hartford basin. When the ice retreated from the north end of Cedar Mountain (Newington-Hartford town line), the Dividend Brook spillway was abandoned and glacial Lake Middletown spread eastward into the southern end of the basin later occupied by glacial Lake Hitchcock. Deltaic deposits (lmw, lme, lmg, and lmwv) as well as lake-bottom deposits (lmb) in the Hartford basin all occur at altitudes accordant with glacial Lake Middletown, but too high to have been controlled by any possible early level of the New Britain spillway. Not until glacial Lake Middletown had lowered to below 110 to 115 ft at the divide (65 ft at The Straits spillway) could the New Britain spillway come into use as the outlet for glacial Lake Hitchcock.

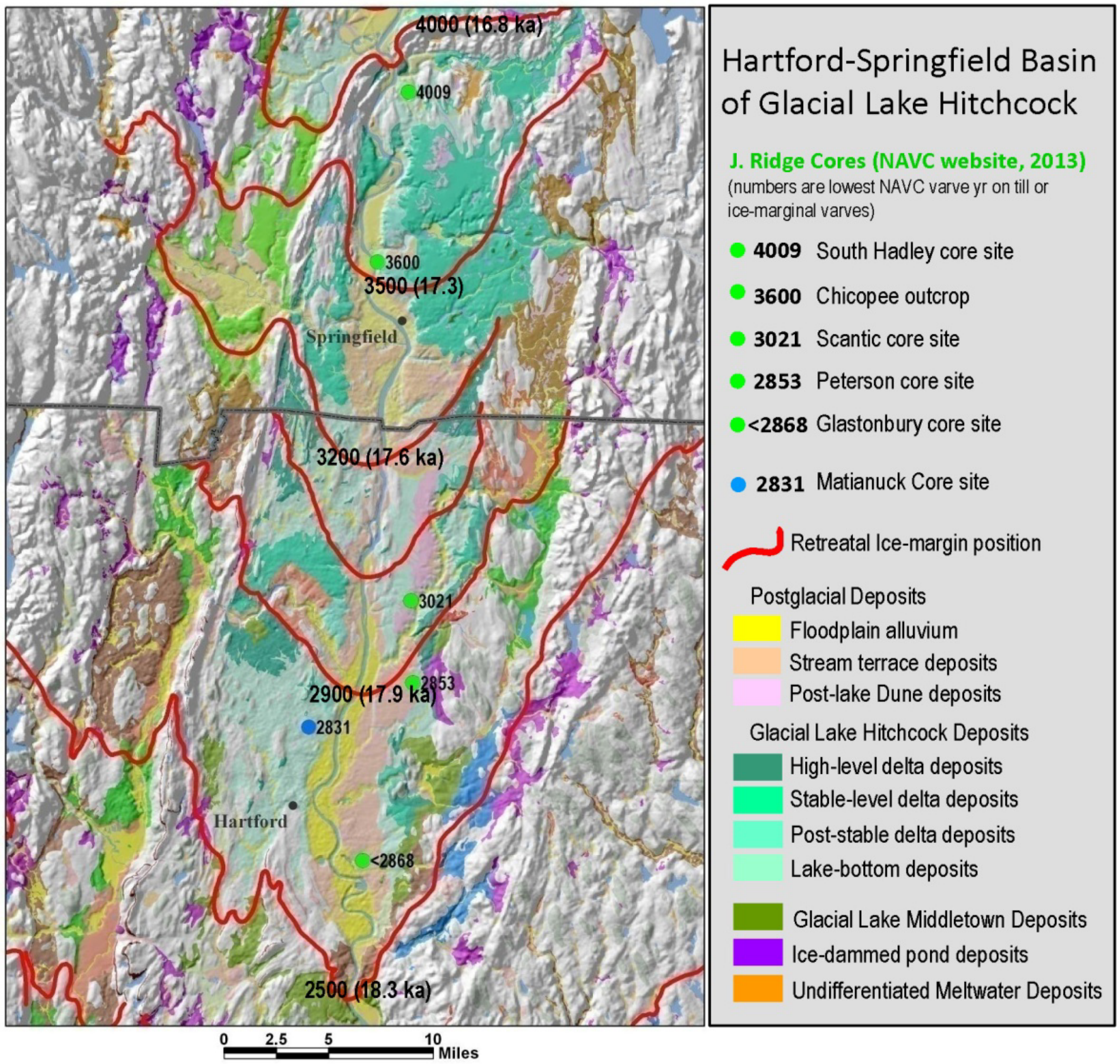


Figure 7. Quaternary geology of the Hartford-Springfield basin of glacial Lake Hitchcock. Numbers associated with ice-margin positions are NAVC varve years (calibrated age in 10^3 years).

Glacial Lake Hitchcock existed in the upper Connecticut River basin in Connecticut, Massachusetts, Vermont, and New Hampshire, lengthening to at least 185 mi as the ice retreated northward to the vicinity of Burke, Vt. The Connecticut River valley was dammed to an altitude of 150 to 160 ft in the vicinity of Rocky Hill and Glastonbury by deposits of glacial Lake Middletown. These deposits and other sediment-dammed pond deposits are often referred to as “the Rocky Hill dam”. The spillway for glacial Lake Hitchcock was not over the dam, however, but at the lowest place across the Mattabesset River drainage divide between the Hartford basin and the Middletown-Berlin basin in New Britain. Glacial Lake Middletown water initially covered the New Britain spillway location and early ice-marginal deltas in the Hartford basin were controlled by glacial Lake Middletown. Not until glacial Lake Middletown had dropped to below 115 ft could the New Britain spillway area emerge and glacial Lake Hitchcock exist as a separate water body; this occurred at about the time that the ice margin was at Windsor and East Windsor (ice-margin positon at varve year 2900 in Fig.7).

During the early life of glacial Lake Hitchcock, the New Britain spillway was eroded into till and older meltwater deposits so that water levels at the spillway dropped from about 115 ft down to 82 ft in altitude. In Connecticut, all ice-marginal and distal-meltwater-fed deltas, as well as one small delta built by meteoric water, record lake levels higher than the longer lived stable level. These deltas show a gradual lowering of the lake level as the ice retreated northward and the New Britain spillway was incised down to bedrock. The bedrock floor of the spillway is at 58 ft in altitude today, thus a 24 ft-depth of water flowing through the spillway was required to produce the 82-ft level. This early phase of glacial Lake Hitchcock is recorded by ice-marginal deltas that are found well into southern Massachusetts and that were built to lake levels between 85 and 95 ft at the spillway. This higher-than-stable-level phase of the lake is referred to as the “Connecticut Phase” (Koteff and others, 1988). Delta levels in Massachusetts indicate that a stable lake level of 82 ft in altitude had been reached by the time the ice margin had retreated to just north of the Chicopee River valley. Deltas that were not associated with the ice margin, but rather were built by meteoric water in most river valleys that entered the lake, also project to the stable lake level. In Connecticut, these include deltas built by the Hockanum River, and the Scantic River, as well as the Farmington River which constructed a large delta northeastward into the lake in the area now surrounding Bradley International Airport. The Bradley Airport delta covers about 20 mi^2 and its entire surface (which is tilted N 21°W in the amount of 4.74 ft/mi) is graded to the stable 82-ft level, indicating that the lake was not affected by glacio-isostatic tilting until after all of its deltas had been constructed in the Hartford-Springfield basin.

An approximate 4,000-year life span for glacial Lake Hitchcock was indicated by Antevs (1922, 1928) through a method of correlating varves in clay pits from Hartford, Connecticut to the north end of the lake basin in St. Johnsbury, Vermont. This method postulates that the silt-clay varve couplets are annual summer and winter layers and that regional seasonal fluctuations affected the thickness of individual varves over the entire lake basin. Antev (1922, 1928) used varved silts and clays of glacial Lake Hitchcock to construct a New England varve chronology (NEVC-NE years) from varve-year 2700 to 7,000. (Antevs chose to begin the chronology at varve year 2700, as this was his estimate for the number of years of ice retreat south of Lake Hitchcock and north of the terminal moraine). Recently, Dr. John C. Ridge of Tufts University has obtained continuous cores through the entire section of Lake Hitchcock varved clay deposits in many places throughout the lake basin and has shown conclusively that varve correlation not only works, but is an extraordinary chronologic and climatic measurement method (see example curve match Fig. 8). These cores, and a USGS core from the Matianuck Avenue site in Windsor, Ct. (Stone and Ridge, 2009) penetrate the entire varve section and identify the basal (oldest) varve resting on till or bedrock at each site. Many new ^{14}C dates on organic debris found in association with the varves have allowed an excellent calibration of varve years to calendar years and a revised (corrected) chronology has been developed, now called the North American varve chronology (NAVC-AM years) (Ridge, 2004; Ridge, 2013, NAVC website http://eos.tufts.edu/varves).

Using the new chronology, lacustrine deposition at the south end of the lake basin began at about varve-year 2700 which is calibrated at 18.1 cal ka. The oldest measured varve from cores in the southern lake basin (2831AM) is at the Matianuck Avenue site. The early Connecticut phase lasted about 800 years until varve year 3500AM (17.3 cal ka) when the ice margin had retreated northward to Chicopee, Mass. By this time the lake had lowered to the stable level which lasted at least until varve year 5200AM (15.6 cal ka) (Stone and Stone, 2012) when the ice margin was at Turners Falls, Mass. Delta levels and varve cores from this area indicate that the lake level dropped about 20 ft at this time indicating that the Rocky Hill dam had been breached and the southern part of the lake (south of the Holyoke Range) was drained.

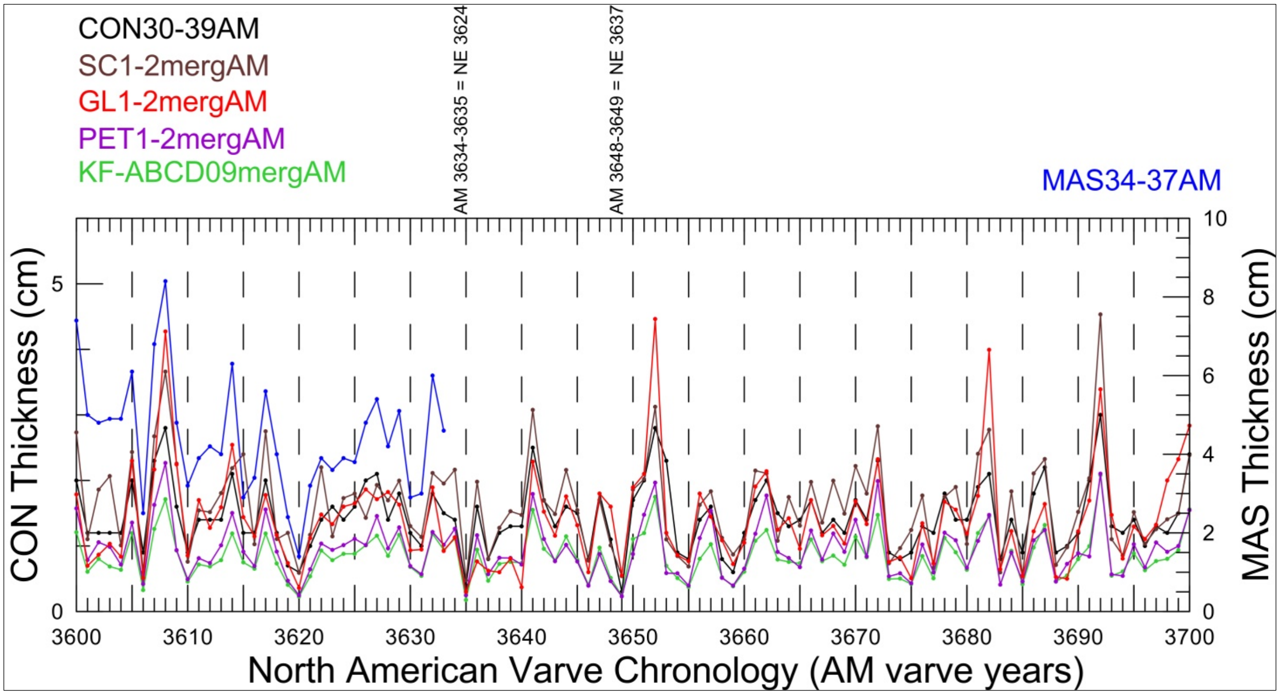


Figure 8. Master plot of varve years NAVC 3600-3700 (Ridge, 2013). Note excellent curve match between Antevs (1928) CON30-39AM Connecticut and MAS34-37AM Massachusetts curves and Ridge cores at Kelsey-Ferguson clay pit, East Windsor (KF-ABCD09mergAM) Peterson Farm core (PET1-2mergAM), Glastonbury core (GL1-2mergAM), and Scantic core (SC1-2mergAM)

The dam was breached most likely by headward erosion of streams on its south side, possibly by ground-water sapping and possibly aided by earthquakes generated by the initiation of postglacial rebound. Regardless of the mechanism by which the dam was breached, glacial Lake Hitchcock could not lower below stable level, much less drain, until its bed was raised by glacio-isostatic tilting. Dam breaching and initiation of isostatic rebound was required in order to establish the lower water-level altitudes recorded in the post-stable phase Farmington River deltaic deposits. Once this process began, it proceeded rapidly as the dam was incised from just above 60 ft in altitude (the stable level at the dam) to just above 40 ft; once this 20 ft of lowering was accomplished, glacial Lake Hitchcock, south of the Holyoke Range, was entirely drained. The newly formed Connecticut River began to incise the drained lake floor along the stream terraces over the 50 mi stretch between the Holyoke Range and the breached dam. Glacial Lake Hitchcock continued to exist north of the Holyoke Range and in the Connecticut River valley in Vermont and New Hampshire at near its original stable level for another 2000 years.

Figure 9 illustrates the point that when altitudes of glacial Lake Hitchcock deltas spillway, and dam are restored to their glacio-isostatically depressed positions (lower part of diagram) before rebound took place, there was virtually no stream gradient down the Connecticut River to Long Island Sound (LIS). In fact, the level of Lake Hitchcock was being controlled by the position of sea level in LIS. Hence, the New Britain spillway could not have been eroded lower than the 82-ft level until glacio-isostatic tilting began.

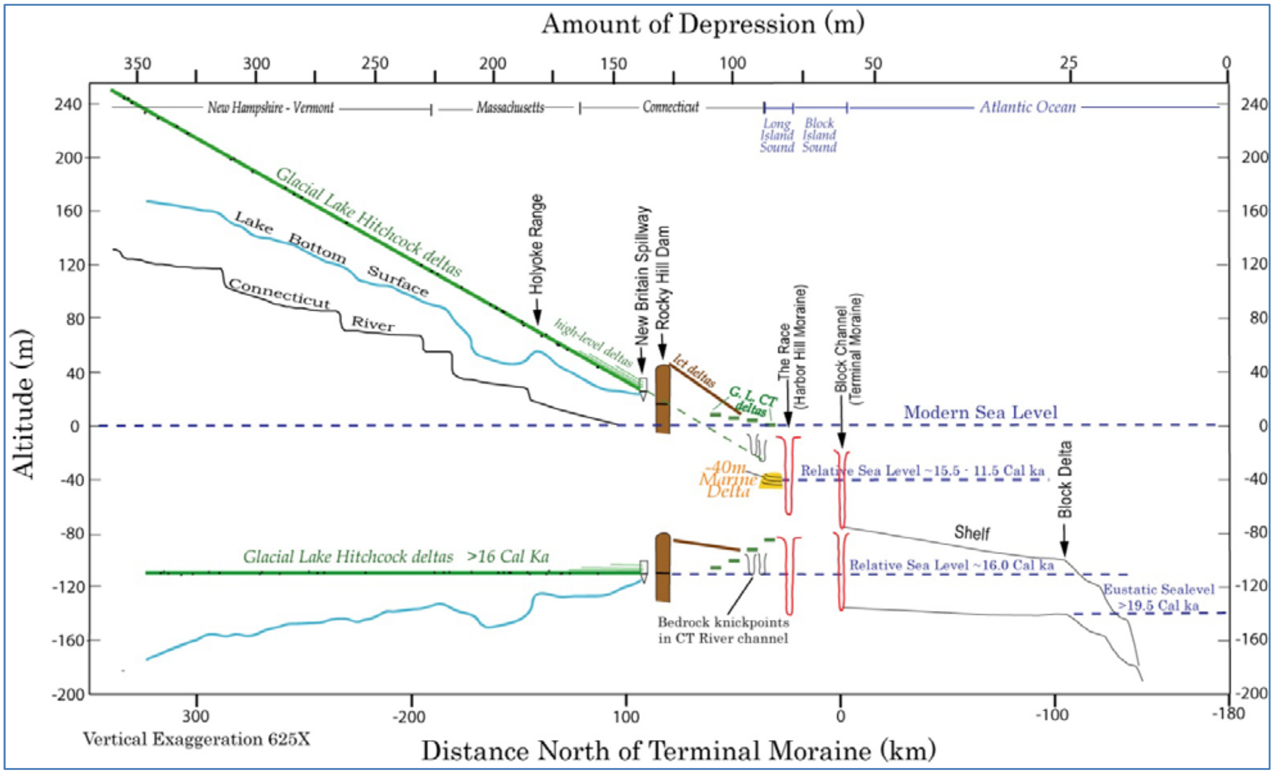


Figure 9. Upper part of diagram shows altitudinal positions of glacial lake deltas (black dots) in the Connecticut valley and other important features to the south as they stand today. Lower part of diagram shows positions of these features when they were glacio-isostatically depressed, that is before glacial rebound occurred.

POSTGLACIAL DEPOSITS AND FEATURES

Postglacial deposits on land in Connecticut include stream-terrace, talus, dune, flood-plain alluvium, swamp, salt-marsh, and beach. Offshore beneath Long Island Sound, fluvial-estuarine channel-fill and marine delta deposits are present. The onset of postglacial conditions was time-transgressive and began several thousand years earlier in the southern part of the State than in the northern parts. In most of mainland Connecticut, postglacial activity consisted predominantly of incision of glacial deposits by meteoric streams along stream-terrace surfaces, followed by the establishment of floodplains at modern levels. Streams had eroded to modern flood-plain levels relatively early, in some cases before 12.0 ka (14.0 cal ka) (O'Leary, 1975; Stone and Randall, 1978). Postglacial winds were intense and widespread as indicated by the ubiquitous blanket of eolian sand and silt that overlies glacial sediments throughout the State and in which the modern soil is developed. The postglacial climate was severely cold for several thousand years following deglaciation. Paleobotanical studies reveal that treeless, tundra vegetation dominated by dwarf willow (*Salix herbacia*), sedges (*Cyprus*, *Carix*), and herbs and shrubs (*Dryas*, *Artemesia*), dated from earlier than 18 ka to about 15 ka, was present in the area (Davis and others, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987; Thorson and Webb, 1991).

Wedge-shaped features with a polygonal ground pattern (Fig. 10), interpreted as ice-wedge casts, deform eolian-sand-capped glacial sediments in numerous localities in Connecticut (Schafer and Hartshorn, 1965; Schafer, 1968; O'Leary, 1975; Stone and Ashley, 1992). These features indicate that permafrost existed locally in areas where substrate conditions were favorable to its formation. The presence of permafrost structures indicates that mean annual temperatures were below 0°C during the early postglacial time interval.

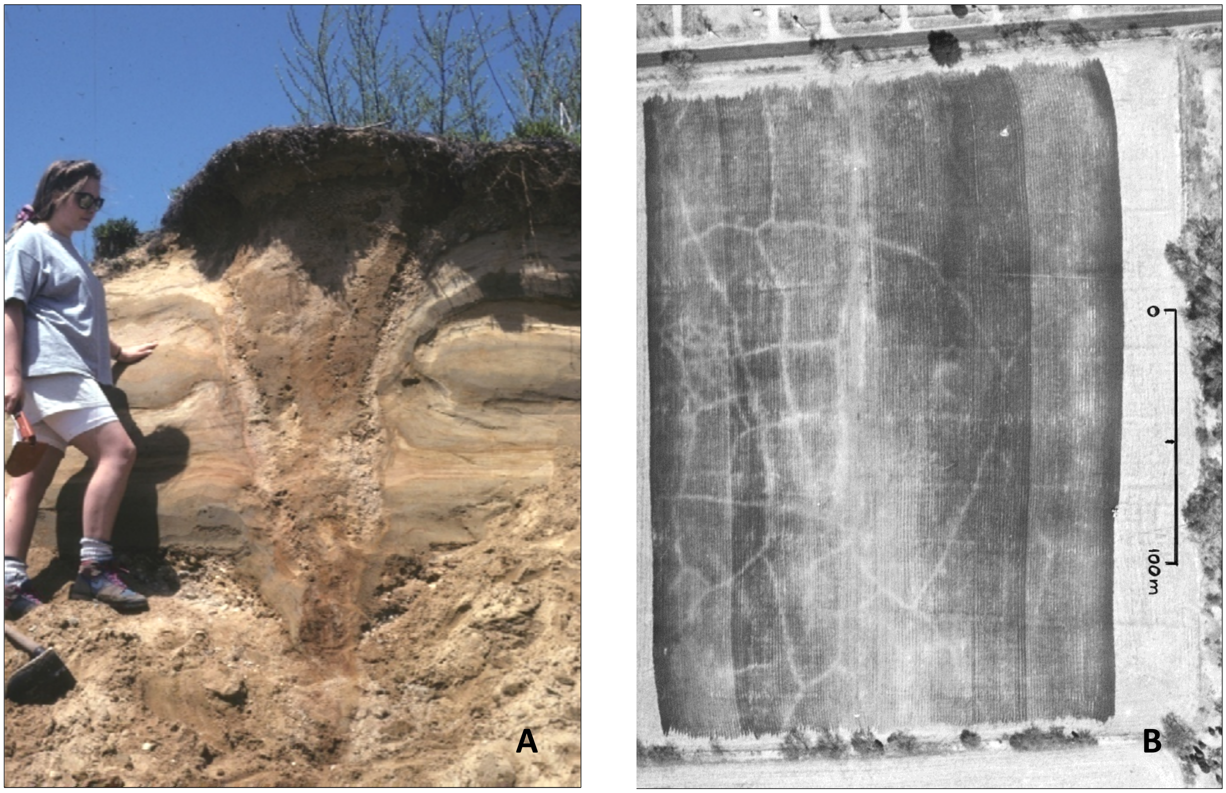


Figure 10. Permafrost features at the Hain Bros. Gravel pit in North Windham, CT. A. cross-section view of ice-wedge cast exposed in pit face. B. Polygonal ground pattern of ice-wedge features seen on aerial photograph of field just north of the Hain Bros. pit.

In the upper Connecticut basin, postglacial conditions were dominated by the continued existence of glacial Lake Hitchcock several thousand years after the ice margin retreated from the area. Extensive fields of eolian sand dunes formed in the treeless environment, indicating the continued effects of strong winds. Dunes are present on the relict deltaic and lake-bottom surfaces of glacial Lake Hitchcock. Dunes on deltaic and high-level lake-bottom surfaces were formed by north to north-northeasterly paleowinds; these surfaces were available as early as 17.3 ka. Dunes on stable-level lake-bottom surfaces were formed by northwesterly paleowinds; these surfaces became available at about 15.6 ka as glacial Lake Hitchcock drained (Fig 11). The presence of hundreds of circular to subcircular, rimmed depressions (interpreted as pingo scars) developed on drained lakebed surfaces provides evidence that severely cold temperatures persisted past the time of glacial Lake Hitchcock drainage (Stone and Ashley, 1989; Stone and others, 1991; Stone and Ashley, 1992). These subtle features are not easily seen on traditional 10-ft contour topographic maps, because their relief is generally less than 3 ft. New Lidar-generated digital elevation models (Figs. 11 and 12) provide excellent illustration of these features.

Paleobotanical records indicate a warming of the postglacial climate at about 14.5 ka, accompanied by reforestation of the landscape by successive spruce, pine, and hardwood forests from 14.5 to 11 ka (Davis and others, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987).

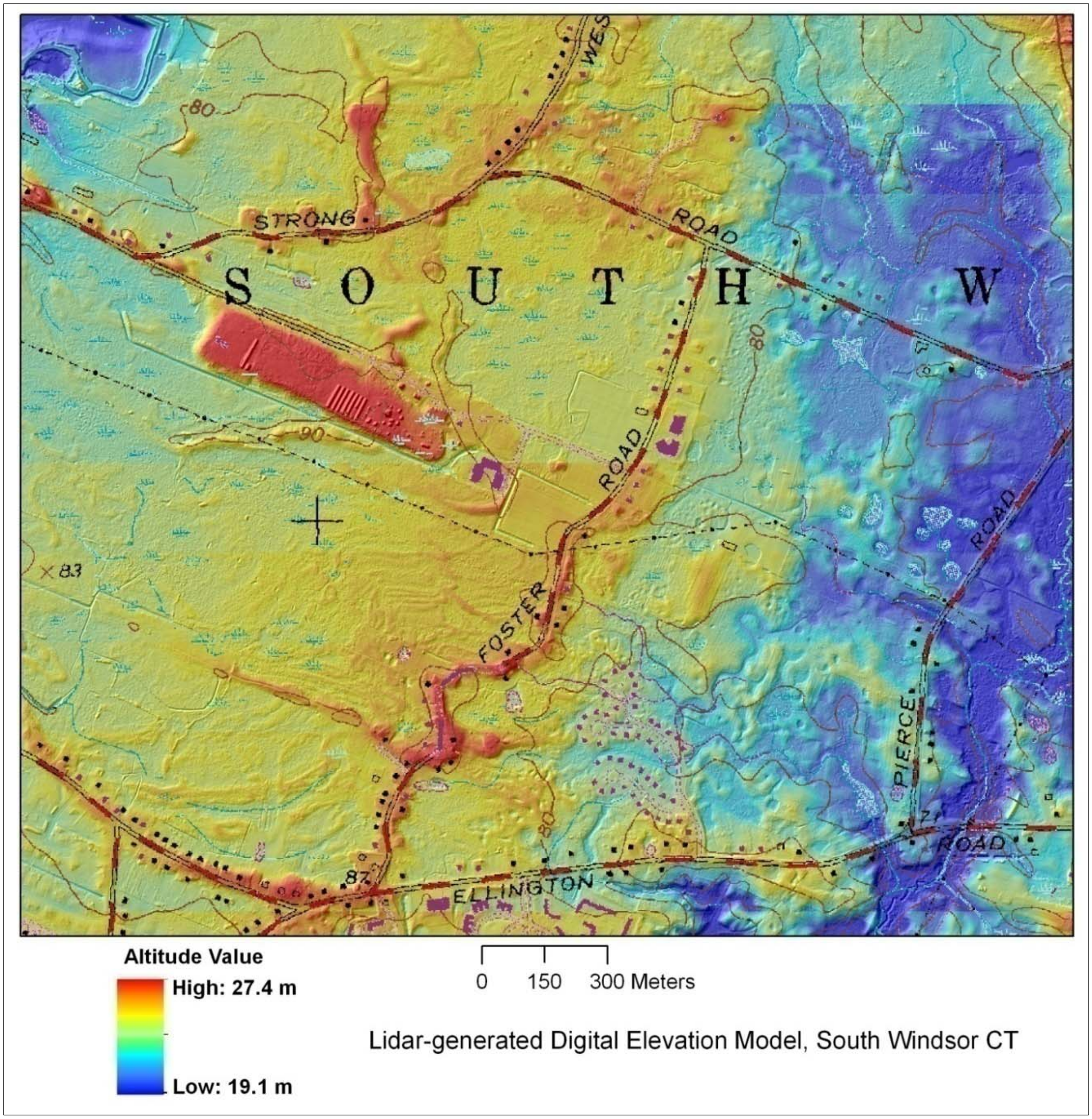


Figure 11. Sand dunes (on higher surfaces in red-orange colors to west) and pingo scars (on lower yellow-blue colors to east) developed on glacial Lake Hitchcock lake-bottom surface in South Windsor, CT. Large rectangular feature is a landfill that interrupts an arcuate dune form.

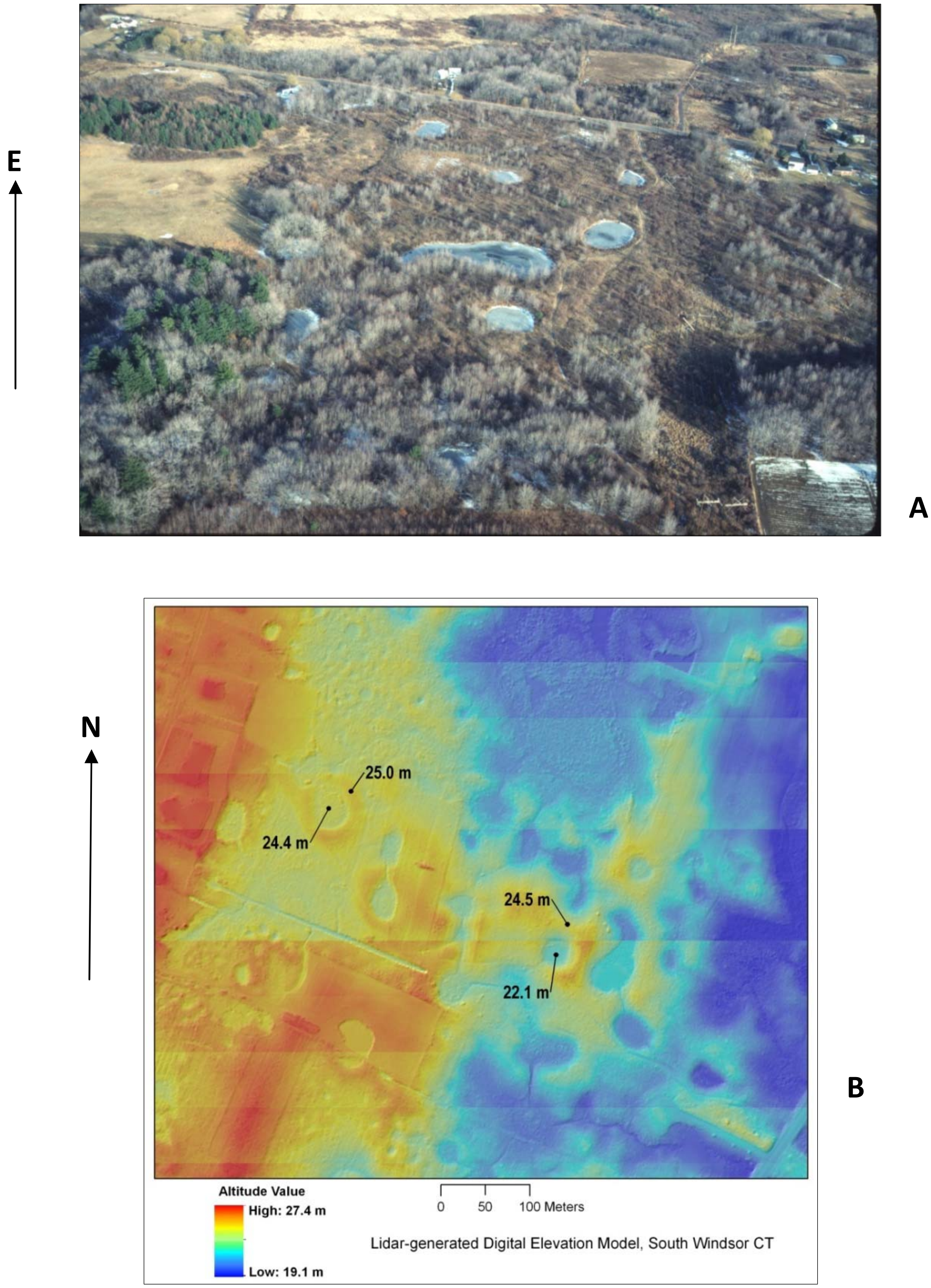


Figure 12. Closer view of pingo scars in South Windsor, CT developed on drained lake-bottom surfaces of glacial Lake Hitchcock. **A**. View from small plane looking east; circular features with ponded water are 30-40 m in diameter. Note circular patterns defined by trees marking rims of depressions in forested areas. Road beyond the ponds is Pierce Road (see Fig. 11). **B**. Lidar image of same area showing low relief of only 1-2 m between center of depression and top of rims.

Fieldtrip Stops

Assembly Point. Dinosaur State Park, Rocky Hill, Connecticut

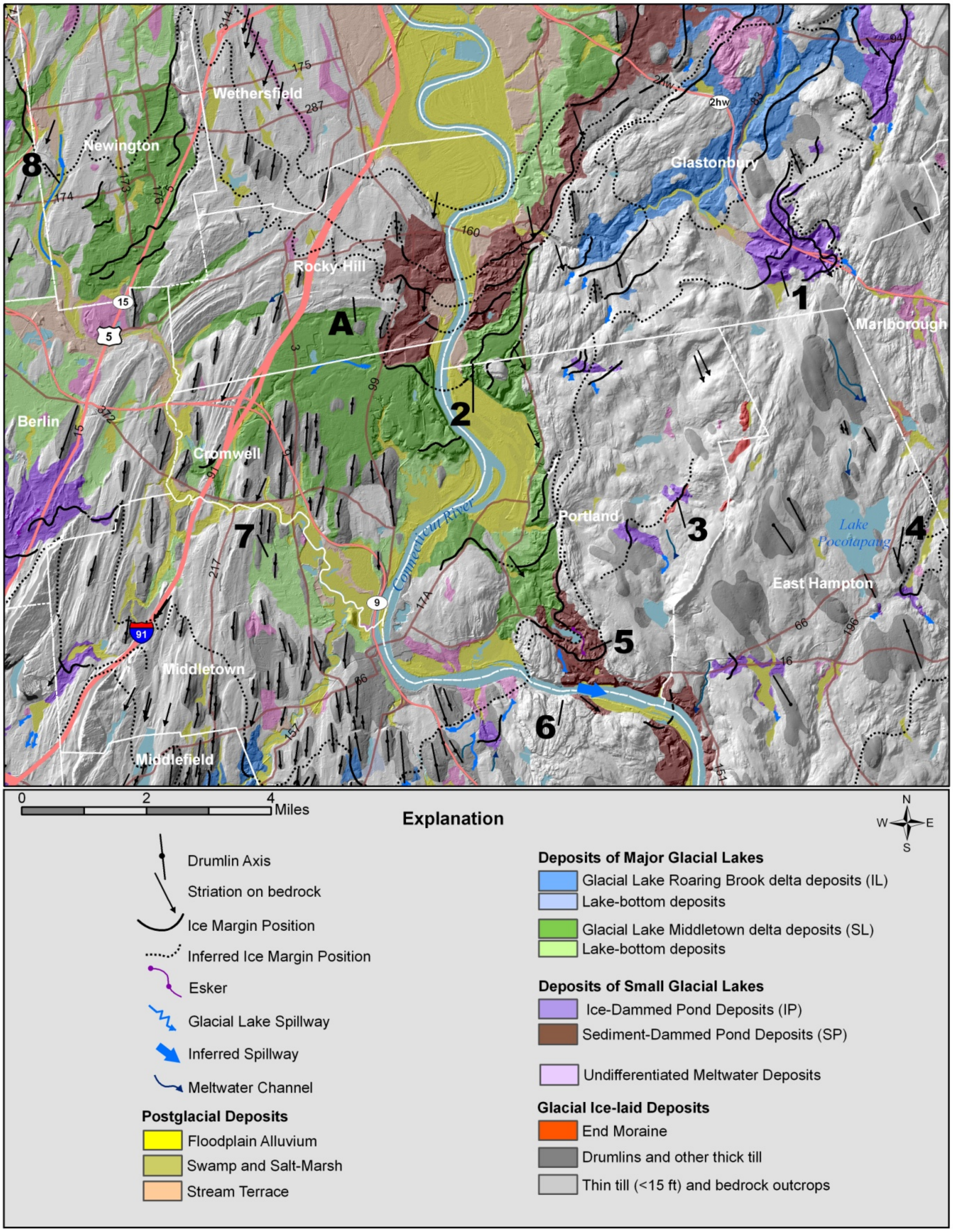


Figure 13. Quaternary geology in the fieldtrip area (Stone and others, 2005). Numbers show location of Stops 1—8 and Assembly Point (A).

Route from Assembly Point to Stop 1.

0.0 Turn left out of parking lot of Dinosaur State Park onto West Street.
0.5 Route passes through area of thin till deposits. Note outcrops of Hampden basalt may be seen on left about 25 yards up Capital Blvd.
0.7 Turn left at third signal onto entrance ramp to I-91N.
1.5 Continue through thin till deposits. Outcrops of East Berlin Formation on right within the next half mile.
3.5 Descend onto stream terrace deposits along the Connecticut River in vicinity of Exit 24.
4.0 View floodplain of Connecticut River to the right. This area is often inundated during normal flood events.
5.2 Take exit 25 for Putnam Bridge, Route 3 North. Cross Connecticut River floodplain surfaces at 5-15 ft in altitude.
6.7 Rise onto stream terrace deposits at 25-35 ft in altitude.
7.5 Exit right onto Route 2 East. Continue over Connecticut River stream terrace deposits incised into glacial lake-bottom surfaces.
8.7 Rise up onto lake-bottom deposits.
9.8 Rising higher in landscape into glacial till deposits. Cross eastern border fault of Mesozoic Basin and enter the Eastern Highlands. Bedrock outcrops are now metamorphic rocks (Glastonbury Gneiss).
11.6 Descend onto deltaic surfaces of glacial Lake Roaring Brook
12.3 Take exit 10 for Country Club Road.
12.4 Turn right on Route 83, Manchester Road.
12.5 Turn left onto New London Turnpike Road.
13.1 Turn right onto Country Club Road.
13.6 Turn left on Mott Hill Road.
14.3 Drive onto collapsed ice-marginal deposits of glacial Lake Dickinson.
14.5 Bear left on Dickinson Road.
14.7 Note steep 70-ft bank on left is frontal foreset slope of ice-marginal delta.
14.9 Turn left into entrance for Glastonbury Earth Products Sand Pit. STOP 1. *NOTE: this is private property; permission from owners is required before entering this site.*

Stop 1. Ice-dammed pond deposits in Dark Hollow Brook valley, Glastonbury CT

Figure 14. Exposure in Glastonbury Earth Products sand pit shows thin delta topset beds overlying thick southerly dipping sandy foresets.

An active sand and gravel pit (Fig. 14) owned by Glastonbury Earth Products off Dickinson Road reveals an excellent exposure of the internal structure of an ice-marginal delta built into a small glacial lake in the WNW sloping Dark Hollow Brook valley (glacial Lake Dickinson of Langer, 1977) (Fig. 15). This deposit is a good example of sediments laid down in an ice-dammed pond depositional system (IP). As the Connecticut valley ice lobe margin retreated westward from the drainage divide, a small lake was impounded in the valley with a spillway at the lowest point across the divide over ledges of Littleton Schist (Rodgers, 1985) at 505 ft altitude. Today the westbound lane of State Route 2 passes through this gap making it difficult to see the original nature of the spillway.

The pit is cut into the southern lobe of an ice-marginal delta with a top surface at 515 ft in altitude. The western and northern slopes of the delta were in contact with the ice margin during its construction, whereas the eastern and southern slopes prograded out into the small lake. The pit exposes about 40 ft of deltaic topset and foreset beds and a 6-ft lower section of finer bottomset sands. In the upper section, about 5 ft of flat-lying gravelly topset beds unconformably overlie about 30 ft of dipping sandy foreset beds (Fig. 14).

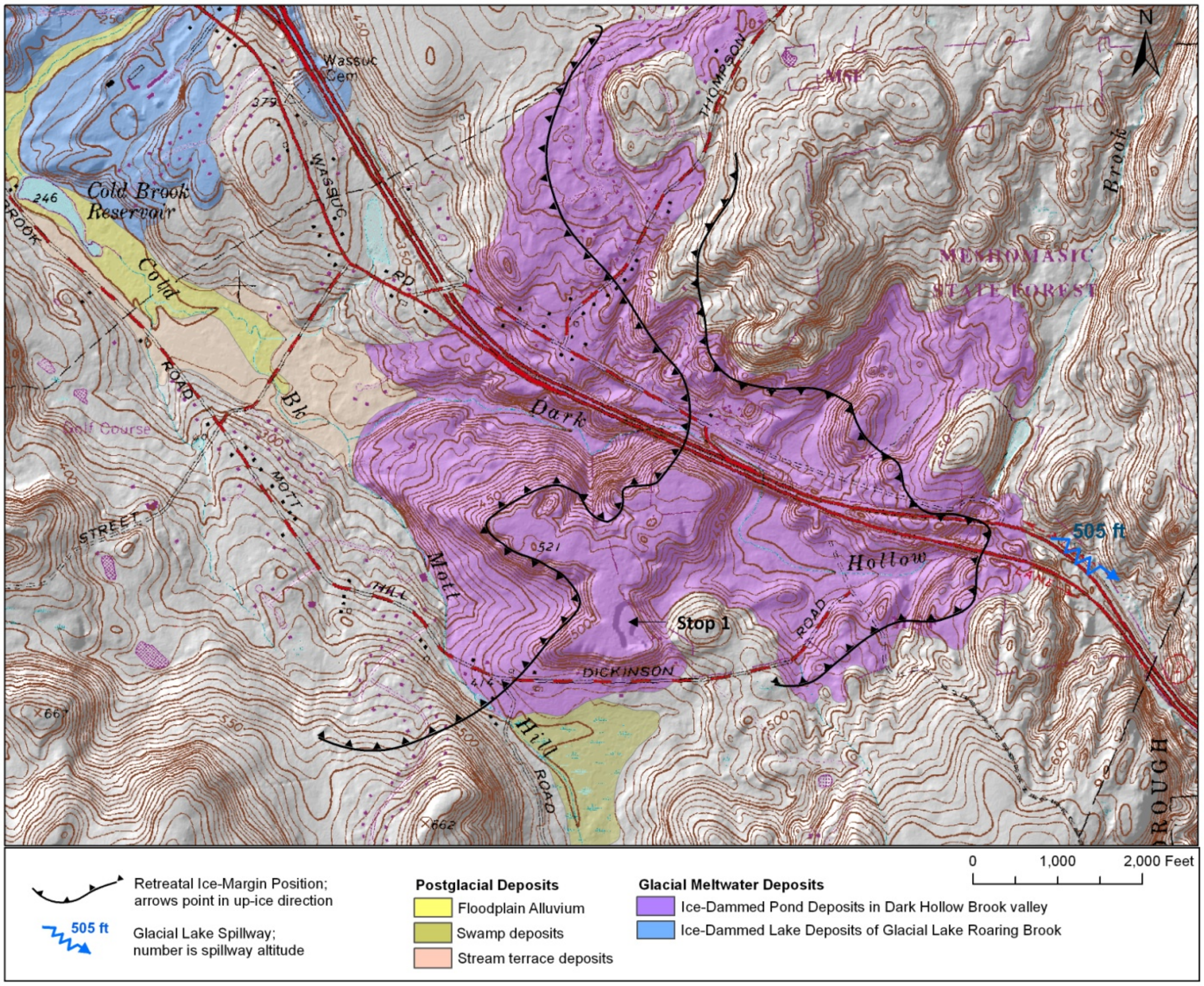


Figure 15. Surficial Geology modified from Stone and others, 2005 in the vicinity of Stop 1.

Mileage **Enroute from Stop 1 to Stop 2.**

0.0 Exit pit access road and turn right on Dickinson Road.
0.4 Turn right on Mott Hill Road.
1.3 Turn left on Country Club Road.
2.5 Turn left on Woodland Street.
2.9 Bear right on Matson Hill Road.
3.2 Cross stream and rise onto small delta deposit of glacial Lake Roaring Brook with surface altitude at 415-420 ft. The 415-ft (highest) spillway for this ice-dammed lake is just south of the upcoming sharp turn to right.

3.6 Matson Hill Road makes sharp turn to right, spillway to the left.
3.8 Note outcrops of Glastonbury Gneiss on left (west) side of road; orchard covered drumlin hill sits higher in altitude above outcrops.
4.1 Turn left on Foote Road.
4.3 To the right deltaic surfaces of glacial Lake Roaring Brook delta deposits can be seen at slightly lower than road level at 280-290 ft in altitude; this delta is graded to a 285-ft spillway across the drainage divide to the left at the upcoming curve to the right.
4.4 Road curves to right, spillway on left.
4.9 Road traverses area of thin till deposits; bedrock outcrops are now Collins Hill Formation
5.3 As slope flattens, cross Mesozoic eastern border fault which here lies beneath glacial meltwater deposits.
5.4 Turn left (south) on Rt. 17. From here to next turn, we are traveling along (to the right) ice-marginal fluviodeltaic deposits of glacial Lake Middletown.
6.4 Turn right on Old Maids Lane at the apple sign, note metamorphic bedrock outcrop on right side of Rt. 17 at the turn. We are traveling across the 175-ft surface of glacial Lake Middletown delta plain (Fig. 13).
7.0 Turn left onto gravel road (town of Glastonbury signs on right). Park and assemble for view to the west across Connecticut River to the other side of the Rocky Hill dam of glacial Lake Hitchcock. Note that depending on access conditions on the day of the trip we may continue to STOP 2 on foot from this spot *OR* we may continue with the driving directions listed below. Turn cars around.
7.2 Turn left on Old Maids Lane.
7.5 Descend onto 50-ft terrace cut through Rocky Hill dam.
7.8 Turn left on Tryon Street.
7.9 Turn left into Glastonbury Bulky Waste Disposal Facility for STOP 2a.

STOP 2a. Excavation at Town of Glastonbury bulky waste disposal site, Glastonbury.

Figure 16. Gravelly, southwesterly dipping deltaic foreset beds exposed in the Glastonbury Bulky Waste Site excavation.

The excavation (Fig. 16) is in an ice-marginal delta in the Cromwell deltaic deposits of glacial Lake Middletown (Fig. 17). In this part of the river valley, the deposits are a series of ice-marginal deltas with surfaces at 165 to 185 ft; a topset-foreset contact at 149 ft was measured by Langer (1977) in a nearby gravel pit. Deltas in Cromwell have depositional free fronts built into open water in the glacial Lake Middletown basin. Together with Dividend Brook sediment-dammed pond deposits, these deltaic sediments form a massive blockage (at 155-175 ft in altitude) in a narrow part of the Connecticut River valley and constitute the sediment dam for glacial Lake Hitchcock. During the high-level and stable phases of glacial Lake Hitchcock, exiting lake water spilled through the New Britain spillway to the west. However, at about 16 ka, the Rocky Hill dam was breeched as glacio-isostatic rebound began. Remnants of a stream terrace at 50 ft in altitude cut through the dam deposits are preserved on both sides of the river in this vicinity. The 50-ft terrace records the dam breeching event when glacial Lake Hitchcock drained and water began to flow through this section of the Connecticut River valley. This drainage only affected the part of Lake Hitchcock south of the Holyoke Range in Massachusetts.

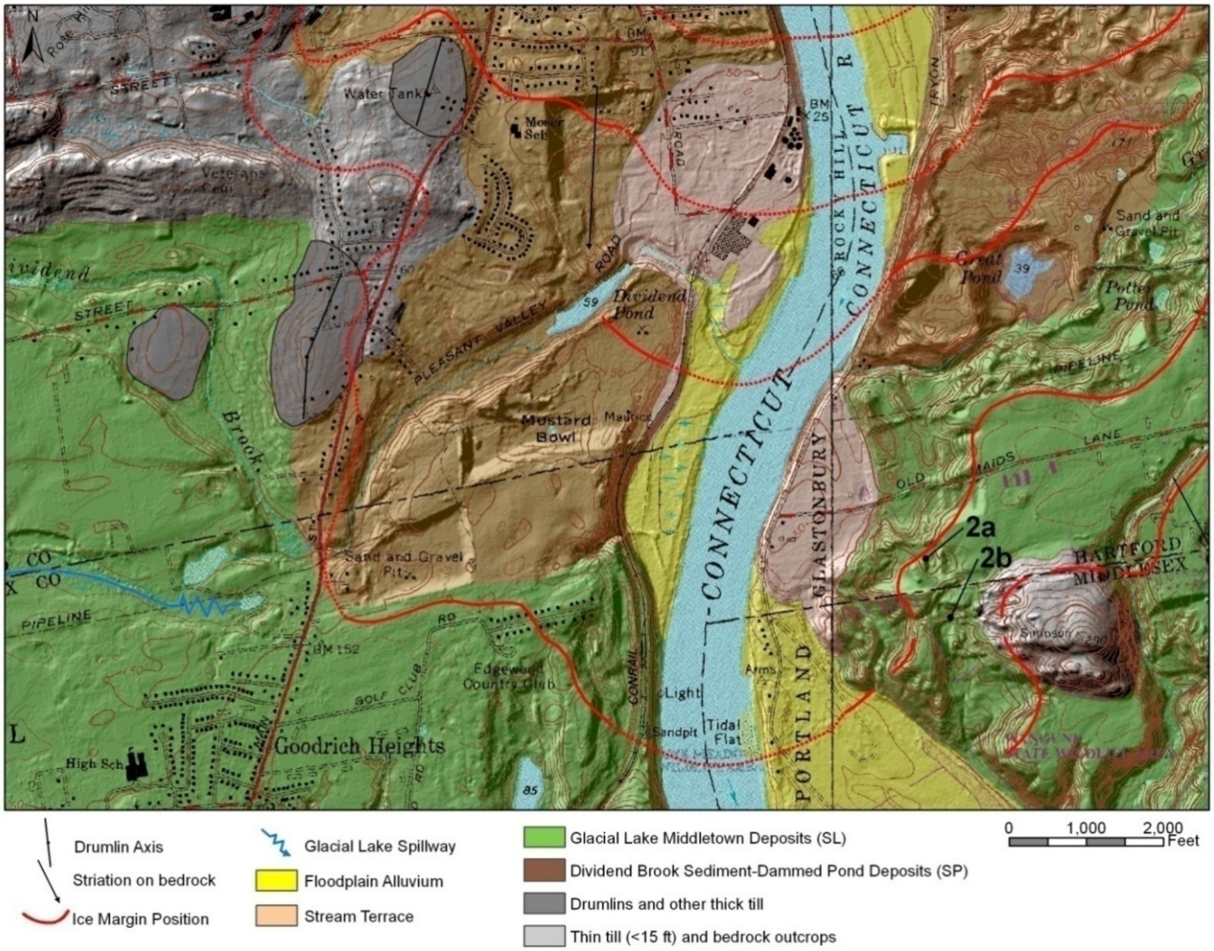


Figure 17. Surficial geology in the vicinity of Stop 2 (Stone and others, 2005). Irregular rectilinear depressions evident in the Lidar DEM are sand and gravel excavation sites.

The relatively flat, noncollapsed delta plain is at 165 ft in this area, but the bulky waste site excavation is cut into the 150-ft surface of an ice-channel ridge at the juncture of walls of a compound kettle (Figs. 17, 19). There are 2 levels of excavation, but only the lower one has fresh faces at this time. In 2005, the upper level pit exposed coarse, ice-proximal sediments, characteristic of the coarse gravel fluvial facies and the sand and gravel ice-channel fluvial facies (appendix 1). Gravel clasts were boulder-sized, subangular to subrounded, with average diameters of 40-50 cm. Beds were collapsed, indicating sediment accumulation in the ice channel between huge ice blocks that later were centers of deep collapse in the compound kettle. In the lower level pit to the west, gravelly foreset beds (Fig. 16) dip 25^{o}-30^{o} southwesterly into the Lake Middletown basin. The strata in this facies consist of alternating beds of coarse pebble sand and pebble-cobble gravel. Coarser beds exhibit open-work textures; bed traces thin and terminate in a down-dip direction in the pit wall. Sandy foresets contain thinly bedded sand and pebble gravel, showing local imbrication of clasts. Below the gravelly foreset beds, finer grained bottomset beds are deformed by folding and faulting caused by melting of buried ice, before deposition of the overlying (undeformed) foreset beds. These upper foresets along the ice-margin position record the last depositional event at the collapsing edge of the delta.

Stop 2b. Striated and meltwater sculpted bedrock outcrop in former Dufford gravel pit.

Figure 18. Outcrop of Portland Arkose exposed in former Dufford gravel pit. A. View is of glacially smoothed north side of outcrop that exhibits southerly trending glacial grooves and striations. B. View of south side of outcrop which is sculpted by meltwater.

A former sand and gravel pit just south of the Glastonbury bulky waste site is no longer in operation, but there remains a large bedrock outcrop of fanglomerate facies Portland Arkose (Fig. 18) that was uncovered during sand and gravel extraction. This outcrop is often visited by geologists interested in older (Mesozoic and Paleozoic) geologic events; but today we will examine the glacial features it also exhibits. The north side of the ESE striking ridge of bedrock (Fig. 19) is glacially smoothed and striated, and also exhibits larger grooves and crescentic gouges indicating southerly ice movement. The south side of the ridge has been sculpted by meltwater that deposited the overlying sand and gravel (now removed).

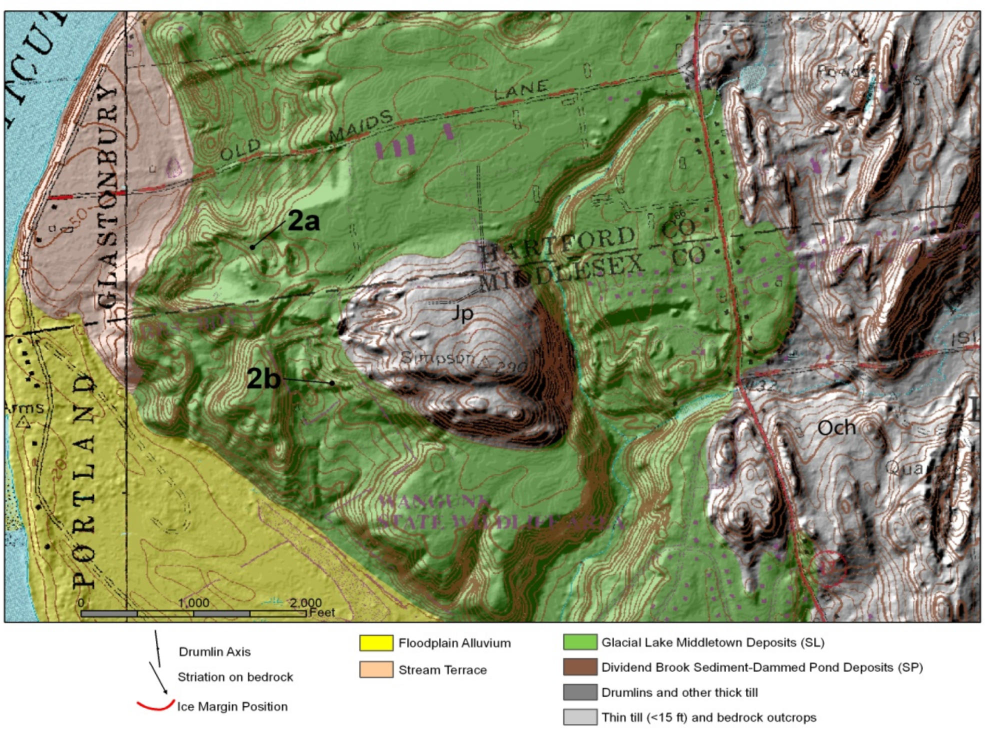


Figure 19. Closer view of Figure 17. Note ESE trending strike ridges of Portland Arkose (Jp) adjacent to Stop 2b evident in the Lidar image (dip to the SSW) and NNW trending strike ridges of metamorphic rocks (Collins Hill Formation Och) to the East. Eastern Border Fault of the Mesozoic basin lies beneath the intervening glacial Lake Middletown deposits.

Mileage **Enroute from Stop 2 to Stop 3.**

0.0 Return to Old Maids Lane. Turn right.
0.7 Note metamorphic bedrock outcrop directly across Route 17 at stop sign. Turn right (south) on Route 17. Continue traveling over glacial Lake Middletown deltaic surface.
1.0 Cross Glastonbury-Portland town line.
1.2 Turn left (east) on Isinglass Hill Road. Enter eastern highlands.
1.5 Cross Hales Brook. A small glacial lake was impounded in the upper reaches of this valley as the ice margin retreated westerly. A series of lowering deltas record water levels at 365 ft, 315 ft, and 275 ft through lowering spillways across the southern drainage divide.
1.9 Lowest 275-ft spillway crosses divide to the right through small subdivision.
2.0 Turn right (south) on Thompson Hill Road.
3.4 Turn left (east) on Old Marlborough Turnpike.
3.8 Turn right (south) on South Road.
4.2 Pass former cranberry bog on left.
4.7 Turn left (east) on Cox Road. Cross Carr Brook. A small glacial lake (IP depositional system) was impounded in this valley as the ice margin retreated westerly. Water was impounded at several lowering levels controlled by spillways across the southern drainage divide from 700 ft in altitude down to 300 ft. Small deltaic deposits record lowering water levels.
4.9 Begin traversing lowest deltaic surface (~300 ft) in the valley.
5.5 Turn right (south) on Woodchoppers Road. Climbing higher in the small lake basin, collapsed ice-marginal deltaic deposits on both sides of road built into a 475-ft water level.
6.2 Park along side of road for hike into Meshomasic State Forest, Stop 3.

Stop 3. End moraine ridges and meltwater-carved channels in Meshomasic State Forest.

Figure 20. A section of bouldery morainal ridge in the Meshomasic State Forest.

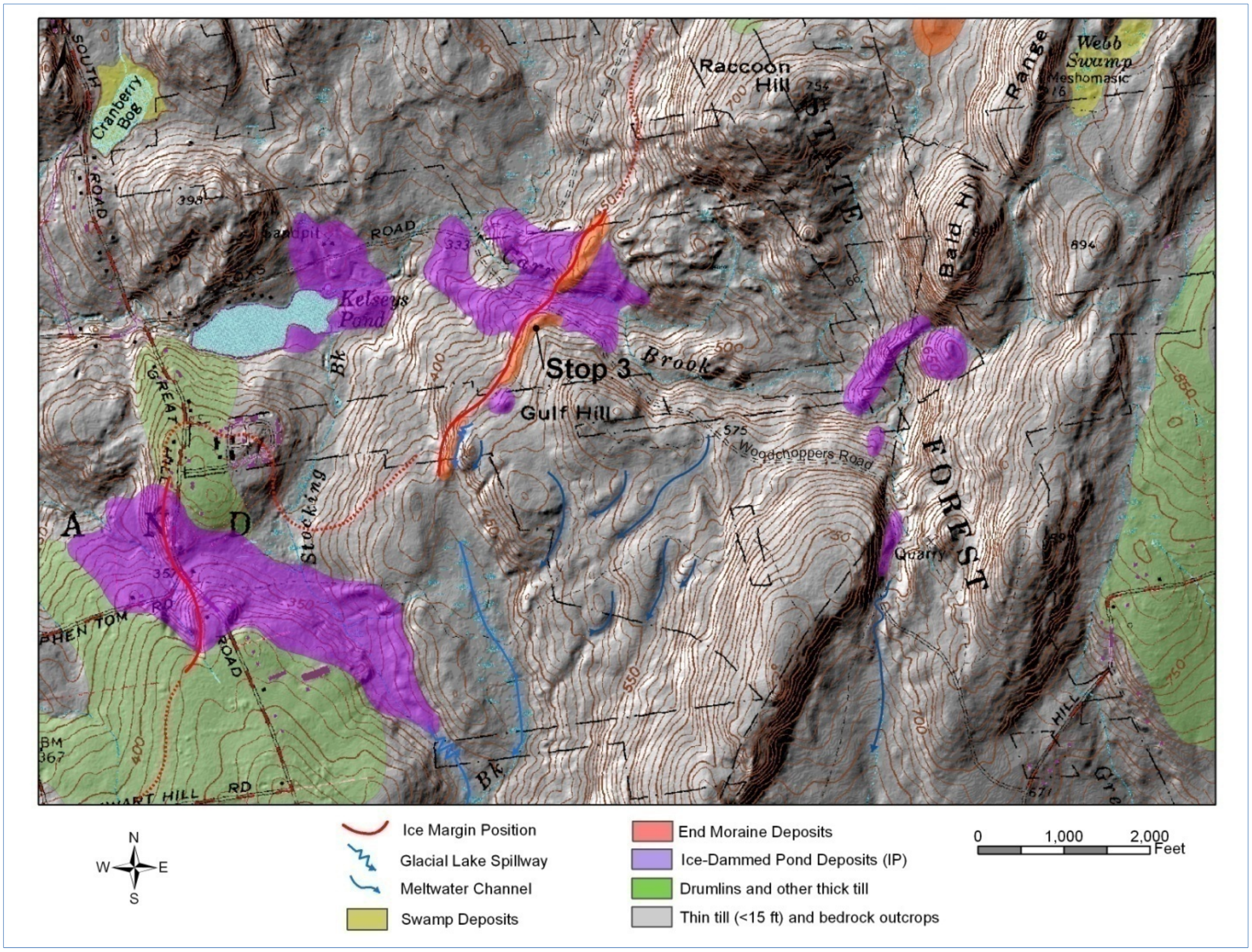


Figure 21. Surficial geology in the vicinity of Stop 3 (modified from Stone and others, 2005).

A discontinuous boulder ridge (Fig. 20) extends along the hillside for about 0.5 miles to the south of Woodchoppers Road. Another short section extends for about 0.25 miles across Carr Brook to the north, where it stands 20 ft higher than surrounding glacial meltwater deposits (Fig. 21). A few other such ridges have been mapped nearby in the Meshomasic State Forest. These bouldery ridges are similar to those prevalent in southeastern Connecticut, but the moraine ridges here do not extend as linear features over long distances. There are no available exposures in this moraine ridge, but other such ridges are composed of sandy ablation till with many surface boulders. This moraine ridge was deposited in association with ice-dammed pond deposits in the Carr Brook valley. As the ice margin retreated westward from the Connecticut River drainage divide formed by the Bald Hill Range, a small glacial lake was impounded in the Carr Brook valley as well as several other westward-sloping valleys. The first and highest spillway for this lake was across the drainage divide through a 705-ft col which we will drive through on the way to the next stop. Deposits graded to this spillway are small and mostly collapsed below the ponding level. A series of lower levels of ponding is recorded by meltwater channels carved into till south of Gulf Hill. Meltwater deposits graded to these channels are scant. The lowest level of ponding at 475 ft was controlled by a spillway between the morainal ridge and the western slope of Gulf Hill. We will hike along the morainal ridge south of the road, and depending on time constaints on the day of the fieldtrip, we may explore various other features in the area.

Mileage **Enroute from Stop 3 to Stop 4.**

0.0 Return to Woodchoppers Road. Continue heading east.
0.4 Road bends right (south) around high bedrock ridge that is Clough Quartzite.
0.5 Abandoned quarry in Littleton Formation across stream to the left.
0.6 Pass by small excavation into pebbly sand on right.
0.7 Road crosses highest spillway (705 ft) across drainage divide for ice-dammed ponding in the Carr Brook valley.
1.2 Turn right (north) on Clark Hill Road.
1.7 Road makes sharp turn to right and begins traversing thick till deposited in the lee of the Bald Hill range.
2.7 Turn right (south) on North Main St. Travel along shore of Lake Pocotopaug.
3.7 Turn left (east) on Route 66, East Main St.
4.1 Hill on right with cemetery is a small drumlin at the south end of the lake.
4.7 Turn right on Lake View Drive and ascend Baker Hill drumlin, Stop 4.
NOTE: this is private property; permission from owners is required before entering this site.

Stop 4. Baker Hill Drumlin, Old till exposure, Route 66, East Hampton, CT

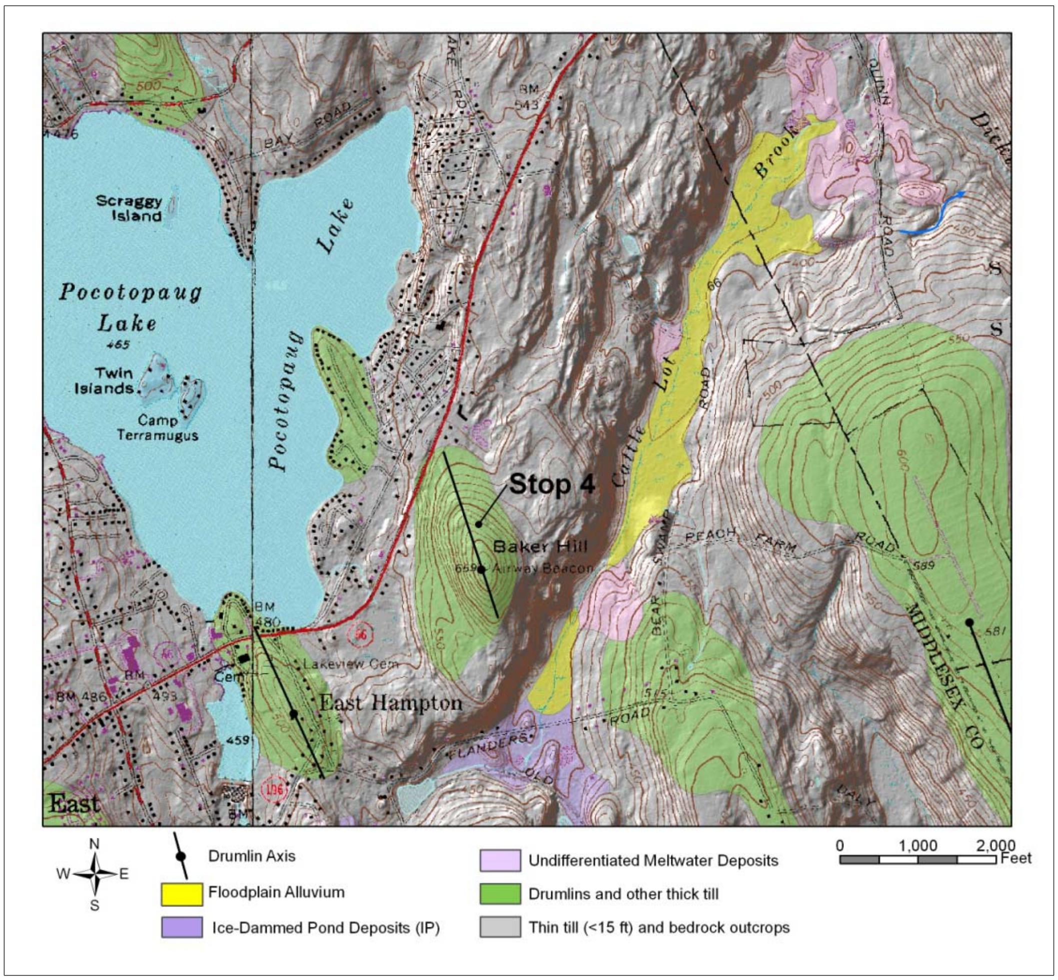


Figure 22. Surficial geology in the vicinity of Stop 4 (modified from Stone and others, 2005).

Baker Hill sits high in the landscape with a summit altitude of 669 ft, atop NNE trending strike ridges of Brimfield Schist (Fig. 22). Till in the drumlin is about 90 ft thick as indicated by the highest surrounding bedrock surface altitude at 580 ft. A former shallow exposure at 540 ft in the lower northwest side of Baker Hill revealed the presence of the lower (old) till in this drumlin. Condominiums now occupy the north slope of Baker Hill and steep scarps deeply incised into the till flank the south sides of each set of condos. These would have been excellent spots to examine the till stratigraphy in this drumlin, but unfortunately the scarps are now landscaped. There is however another scarp higher up the hill at 630-640 ft altitude which has no building at this time. With some shoveling by fieldtrip participants, we can probably dig out small exposures of the two different tills known to compose most drumlins in Connecticut.

Mileage **Enroute from Stop 4 to Stop 5.**

0.0	Descend Baker Hill and turn left (west) on Route 66, East Main St.
3.4	Continue straight at stop light, junction with Route 16. High scarps behind Global Self Storage on left are remnants of former sand pit in 380-ft delta built into small ice-dammed pond in upper reaches of Mine Brook valley.
3.8	Cross Mine Brook.
4.2	Continue straight through stop light at Cobalt.
5.1	Note road cuts through metamorphic bedrock on both sides of road.
5.4	Leave high bedrock area and enter buried former channel of the Connecticut River today filled with glacial meltwater deposits (sediment-dammed pond deposits) (Fig 23). Deltaic surfaces here are at 140-150 ft. Former gravel pit on south side of road revealed very coarse gravel topset beds overlying sandy foresets. Well logs in this vicinity indicate more than a 280-ft depth to bedrock.
5.6	50-ft deep kettle on north side of road marks the thalweg of buried Connecticut River valley.
6.1	Turn sharp right (east) on Middle Haddam Road at stop light.
6.4	Turn left onto Courtney Lane.
6.6	Proceed to end of cul de sac. Park for Stop 5. *NOTE: this is private property; permission from owners is required before entering this site.*

Stop 5. Jobs Pond kettle and Harmon Gravel Pit, Middle Haddam Road, Portland CT

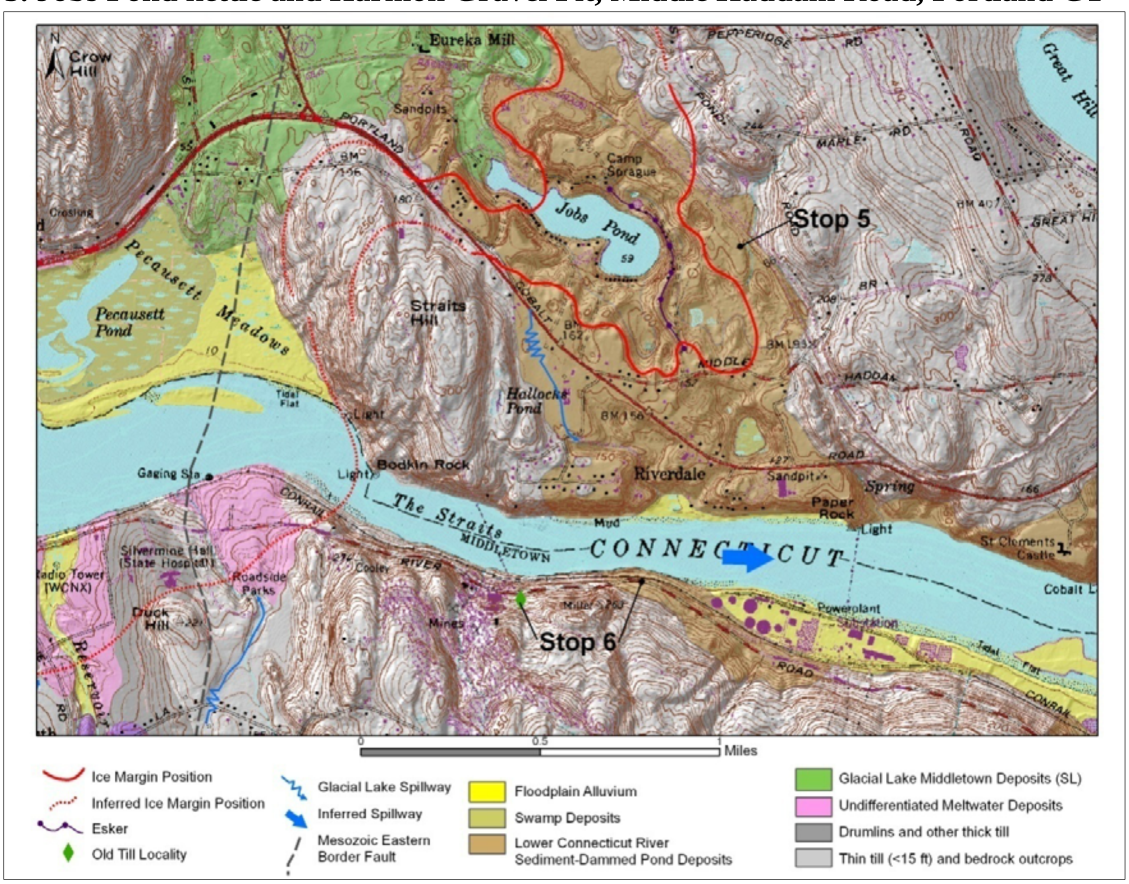


Figure 23. Surficial Geology in the vicinity of Stop 5 and 6 (Stone and others, 2005).

Stop 5 lies within the northern part of the lower Connecticut River sediment-dammed pond deposits. Non-collapsed delta-plain surfaces are at 155 ft in altitude (Fig. 23). These deposits surround Jobs Pond, a scenic, compound-kettle pond that overlies the thalweg of the buried bedrock valley of the Connecticut River. Jobs Pond has a surface altitude shown on the topographic map at 59 ft above sea level, but water levels vary greatly from year to year. Local lore has it that the pond is bottomless; if there was no sand and gravel beneath the ice block that melted to produce the pond, it may be about 180 ft deep as bedrock beneath it lies about 120 ft below sea level.

At least two ice-marginal deltaic morphosequences extend across the buried valley here, including a fine example of an ice-channel feeder (esker) of the delta along Middle Haddam Road. The deposits in this vicinity locally are more than 250 thick and originally extended across the area of the present river channel to the south. These deposits completely filled the valley to 150 ft altitude and formed the northern part of the dam for glacial Lake Middletown, deltas of which, in turn, formed the drift dam (Stop 2) for the long-lived glacial Lake Hitchcock to the north. Following drainage of the glacial lakes in the Connecticut valley and post-glacial uplift, the ancestral Connecticut River was forced to flow on the lower lake-bottom plain of Lake Middletown to the west, rather than along the sediment-filled bedrock valley beneath Jobs Pond. The future course of the river and its local subsequent history was set by the nature of ice-margin retreat and valley-filling lake sedimentation.

We will first climb up onto the esker ridge to the west of the cul-de-sac and walk north along the ridge to view the Jobs Pond kettle; then return to the cul-de-sac and proceed into gravel pit along a farm road to the northeast. The pit exposes 12-15 ft of flat-lying delta topset beds that are red-brown in color. These overlie about 10 ft of pink pebbly sand foreset beds (Fig. 24). The topset-foreset contact is not well exposed in the pit, but it appears to be near the base of the gravel beds. Since the land surface is at about 160 ft here, the topset-foreset contact that records lake level is at about 145 ft in altitude. The spillway control for this deltaic morphosequence was over slightly earlier deltas down valley, but these surfaces have since been removed by Connecticut River incision into the deposits (see inferred spillway symbol, Fig 23).

Figure 24. Topset gravel and foreset sand exposed in the Harmon sand and gravel pit, Portland, CT.

Mileage **Enroute from Stop 5 to Stop 6.**

0.0 Return to cars and proceed south on Courtney Lane.
0.2 Turn right (west) on Middle Haddam Road.
0.3 Turn right (northwest) at stop light on Route 66.
0.5 Note meltwater spillway channel on the left behind gift shop and Arrigoni Winery. This channel was cut across the 155-ft deltaic surface of the earlier deltaic morphosequence by meltwater spilling from the next sediment-dammed pond to the north (the last of the morphosequences in lower Connecticut River sediment-dammed pond deposits.
0.6 Rise slightly off meltwater deposit surface into area of shallow bedrock. Note outcrops of Collins Hill Formation on left (160 ft in altitude). Just a few hundred feet to the east (right side of road), the bedrock surface drops off to greater than -120 ft in altitude.
1.3 Proceed straight continuing on Route 66 at stop light for Route 17 junction. Road has descended down onto 145-ft deltaic surface of early deposits of glacial Lake Middletown. We cross the Eastern Border Fault of the Mesozoic basin here; easternmost Mesozoic bedrock outcrops can be seen a short distance up Route 17.
2.0 Note outcrops of Portland Arkose at base of Crow Hill on right.
3.3 Turn left at stop light onto Main Street, Route 17A.
3.8 Cross Connecticut River via the Arrigoni Bridge. Continue onto Main St., Middletown.
4.3 Turn left (east) at stop light follow signs for Route 9 south.
4.5 Turn right (south) onto Route 9 south.
5.7 Take exit 12 for Silver Street.
5.9 Turn left (east) onto Silver Street at end of ramp.
6.0 Continue straight at Stop light to stay on Silver Street. Rise up and over two drumlin hills. View of Connecticut River to the left (north).
6.7 Bear right onto River Road. Note forested hill behind hospital with green roofed buildings; this is Duck Hill which is the easternmost Mesozoic outcrop in this vicinity.
7.2 Note westernmost metamorphic bedrock outcrop (Collins Hill Formation). Eastern border fault lies buried beneath glacial deposits somewhere in between.
8.4 Turn cars around at NRG power plant entrance road, and drive a short distance back.
8.6 Park along side of road and exit vehicles to view “The Straits” and discuss the former sediment dam for Lake Middletown that once occupied this section of the valley. Then proceed to till cut about 1800 ft farther along River Road (south side).

Stop 6. View of “The Straits” of the Connecticut River and old till locality, River Road, Middletown.

From this vantage point looking toward the left (west) we see “The Straits” where the Connecticut River leaves the Mesozoic Basin and enters the Eastern Highlands though a gorge cut through ledges of Collins Hill Formation (Fig. 23). East of there, the River cuts through the 150-ft surface of glacial meltwater deposits in the Jobs Pond valley that we saw at Stop 5. These deposits were continuous from side to side in the valley as the ice margin was retreating through here. Before incision by the Connecticut River, this sediment provided the dam that impounded glacial Lake Middletown in the basin to the west. We can also see bedrock outcrops drop off and disappear through the gorge beneath the glacial meltwater sediments where the buried

Figure 25. View of The Straits looking north from River Road, Middletown.

Figure 26. Old (Illinoian) till exposed in gully cut south side of River Road in Middletown. Two-ft shovel handle for scale.

ancestral Connecticut River valley joins the present course.

From this view point, proceed about 1800 ft west on River Road (Fig. 23) to examine an exposure of old till in a bank on the south side of the road (Fig. 26). This area has not been mapped as thick till and bedrock outcrops are present nearby. There is, however, a small area where stream runoff from the old feldspar quarries higher up the hillside has carved gullies into till banked onto the lower valley slope. This till outcrop was better exposed in the past and was identified as old (Illinoian) till by J.P. Schafer in his 1978 field notes: "Slumping of roadside cut bank shows fresh washed surface in red-brown old till, 2 m high. Platy jointing shows beautifully on surface—spacing as close as 1-3 mm above, 5-10 or even 15 mm below. Till is very compact and hard, breaks along jointing, which is weakly to moderately stained (with iron-manganese). Sample collected in middle of upper half. Color (SCS, soil moist) is 5YR4/3.5. Jointing dips outward to NNW as much as 30^{o}. Very small stone content, certainly much less than 5%, perhaps only 1-2%."

It is more difficult to distinguish upper and lower till in the red-brown tills because the distinctive olive-gray color of oxidized old till derived from crystalline rocks is not a factor. The color of upper and lower red tills is very similar. A gully higher up the hillside exposes a less dense, sandier, more stony till that may be either the mixed zone containing clasts of lower till, and/or the upper till. Larger boulders are present in this upper unit. The exposure face needs to be cleared off and examined more closely.

Mileage **Enroute from Stop 6 to Stop 7.**

0.0 Proceed west along River Road, retracing route back to Main Street.
1.4 Bear left on Silver Street.
2.1 Proceed straight through Stop at Eastern Drive to stay on Silver Street.
2.2 Cross under Route 9.
2.7 Turn right on Main Street Ext.
3.6 Turn left (west) at stop light onto Washington Street (continuation of Route 66).
3.9 Ascend up onto north ends of drumlin hills beneath Wesleyan University on left.
4.2 Turn right (north) at stop light onto Newfield Street, Route 3.
4.5 Cross Coginchaug River.
4.9 Proceed straight at stop light continuing on Newfield Street. Drumlin hill on right.
5.3 Descend onto lake-bottom surface of glacial Lake Middletown, which lies at 35-45 ft altitude. Abandoned clay pits are a short distance east of the road.
6.0 Turn left (west) on Mile Lane.
6.4 Turn right (north) on Kaplan Drive.
6.8 Park in parking lot for Lawrence School and proceed to Stop 7.

Stop 7. Lake-bottom deposits of glacial Lake Middletown, Newfield section of Middletown.

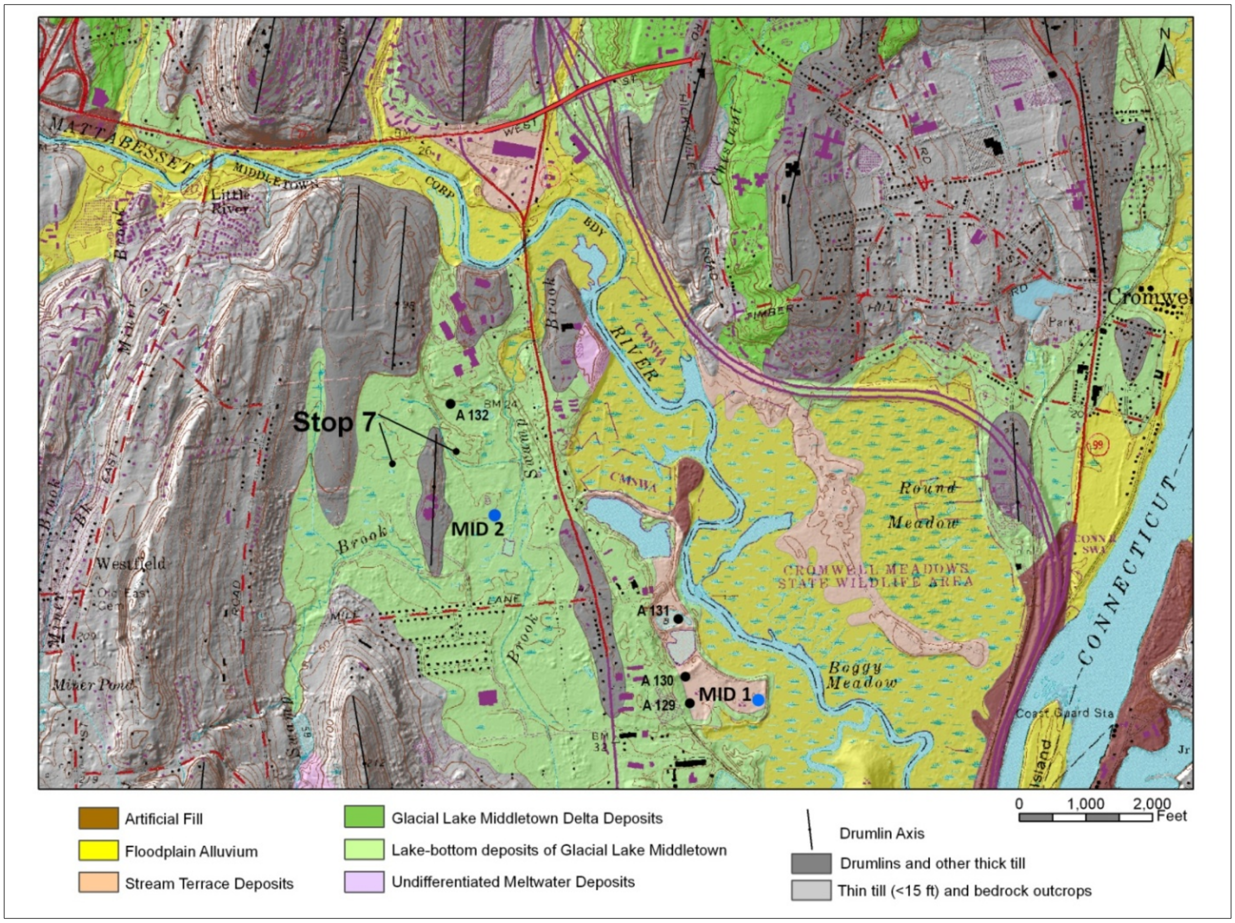
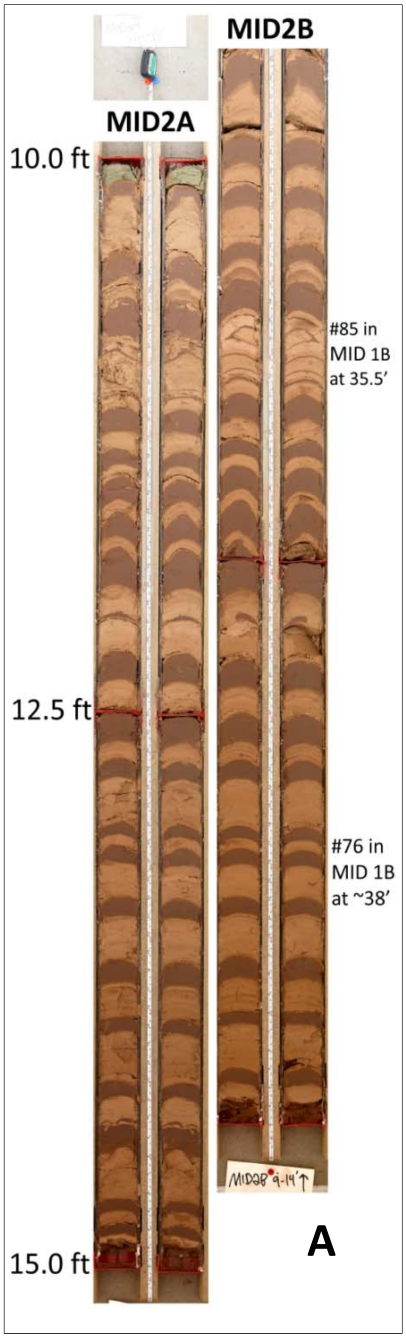


Figure 27. Surficial Geology in the vicinity of Stop 7 (modified from Stone and others, 2005). Blue dots are locations of USGS Cores MID 1 and MID 2. Black dots are locations of Antevs 1928 localities from which he generated his Berlin and Newfield Series curves.

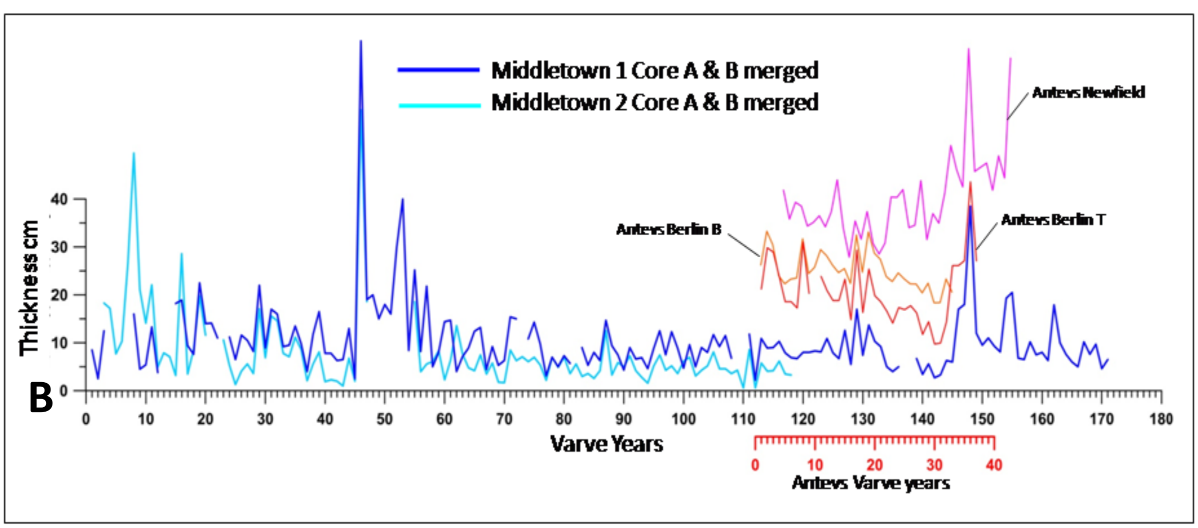
Lawrence School sits slightly higher than the lake-bottom surface on a small drumlin that was not covered by lake-bottom sediments although it is well below the approximate 120-ft lake level of glacial Lake Middletown (Fig.27). Varved lake sediments were presumably deposited by underflow currents descending from delta fronts and accumulated only in the deepest parts of the lake. Varved lake-clay surfaces are preserved at 35-45 ft in altitude where they have not been postglacially eroded by Swamp Brook and its tributaries and the Mattabesset River. The lake clay is thickest near the Mattabesset River. MID1 cores penetrated 75 ft of varved clay sitting on top of dense red till. MID2 cores farther west penetrated 39 ft of clay overlying red till. In the past, excellent exposures of varved sediments were available in the Kane Brick Co. clay pits east of Newfield Street. At the clay pits varved sediments are overlain by post-lake stream-terrace sand and gravel deposits constructed along the Mattabessett River by water exiting the New Britain spillway from glacial Lake Hitchcock to the northwest. Exposures at the Kane clay pits are described in Stone and others (1982) and London (1985). Today there is no active excavation of clay and most of the clay pits have been filled in. We will walk from the school parking lot north a short distance into the woods where the lake-bottom surface is incised by a small stream; a cut in the stream bank reveals a small exposure of varved clay, which we can clear off for a better view. We will also bring along on the fieldtrip some of the core sections discussed below.



In an attempt to extend the New England Varve Chronology southward into the Lake Middletown basin, the USGS obtained side-by-side, vertically offset continuous core at two drill sites (MID1 and MID2, Fig. 27). These cores sampled the entire lake-clay section and penetrated till at the base of the section. A total of 171 varve couplets were measured. Varve plots from the two localities (which are about a mile apart) match well with each other (Fig. 28B). Although no overlap was found with the Lake Hitchcock curves, at least 171 years are recorded by varve deposition in the Lake Middletown basin. The upper part of the curves from the cores match with Antevs' Berlin and Newfield series (Fig. 28B) and add 16 more years to the chronology.

Figure 28A shows a section of the MID 2 side-by-side, vertically offset cores from 10 ft to 15 ft depth. There are 24 varves in this 5-ft section (varve years 68 to 92). These are thick red varves, couplets range in thickness from 1 to 6 inches (average 2.5 in). Despite the fact that the lake clay section is 75 ft thick at the MID 1 site and only 39 ft thick at the MID 2 site, the same years are recorded at both sites indicating that the varves are thicker in the deeper parts of the lake. At both sites, the top of the varve section has been removed. At site MID 1, the top of the section has been eroded by water that deposited the stream terrace deposits; the top of the lake section at the MID 2 site has been postglacially eroded by Swamp Brook and has also been subjected to the deformation that formed the depressions interpreted to be pingo scars.

Figure 28. **A.** MID 2A and B split cores at 10 to 15 ft depth. **B.** Varve plots from Middletown cores 1 and 2 (in blue) and correlation with Antevs curves from localities 129-136 (Antevs, 1928) at the Kane brick pit in Newfield and the Donnelly brick pits in Berlin.



Similar to the lake-bottom surfaces of glacial Lake Hitchcock (Figs. 11, 12) and other glacial lakes in southern New England, those of glacial Lake Middletown also exhibit fields of small rimmed depressions that have been interpreted as pingo scars (Stone and Ashley, 1989; 1992; Stone and others, 1991;) indicative of the presence of cold climate conditions at the time of lake drainage. Glacial Lake Middletown was drained more than 1000 years earlier than the southern basin of Lake Hitchcock, when the ice margin was not far away, making it more probable that the climate was still cold enough to support the development of permafrost. Most likely these features formed as soon as the surface was available because they do not occur on younger stream terrace or alluvial surfaces.

There are many small circular to subcircular vernal pools with subtle raised rims in the wooded area just northwest of the Lawrence School (Fig. 29). Here the features range from 10 ft to about 50 ft in diameter. This particular group has been preserved whereas many others have not escaped suburban development. These features are consistent with pingo-scar permafrost depressions.

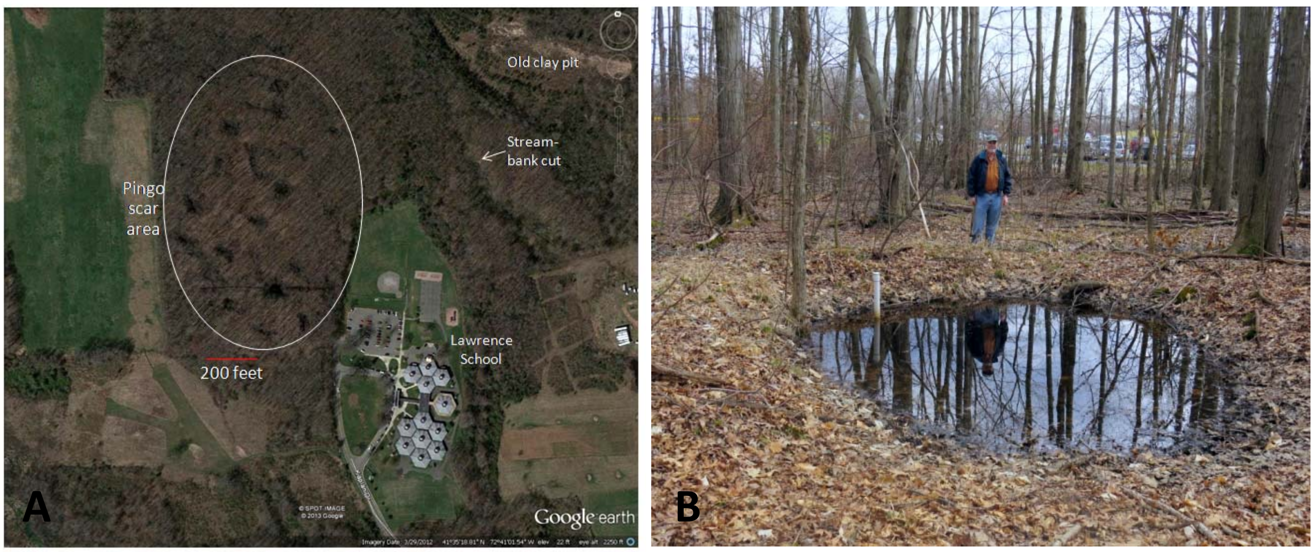


Figure 29. **A.** Google image March 2012; dark, circular areas just northwest of school are small vernal pools in bottoms of pingo-scar depressions. **B.** Photo of rimmed depression in the woods just west of Lawrence School.

Mileage **Enroute from Stop 7 to Stop 8.**

0.0 Leave parking area and retrace route on Kaplan Drive.
0.4 Turn left (east) on Mile Lane.
0.8 Turn left (north) on Newfield Street, Route 3. Traveling over lake-bottom surface.
1.8 Cross Mattabesset River cut into lake-bottom surface. Surrounding wetlands are tidal.
1.9 Ascend to 25-ft stream-terrace surface.
2.2 Stop light at Route 372. Continue straight to stay on Route 3, now Shunpike Road.
2.5 Overpass for I-91. We have entered into a drumlin field; note high hill on left.
2.8 Another drumlin on the left. There are 11 drumlins in this 2 mi^2 area.
3.8 Emerge onto 135-145-ft Cromwell delta surface of glacial Lake Middletown.
4.0 Descend into swale crossing the road that is the distal end of the Dividend Brook spillway that cuts the Cromwell deltaic surface and served as the spillway for Dividend Brook deposits (Fig. 13, 17).
4.2 Cross Cromwell-Rocky Hill town line.
5.0 Cross over I-91 again. Entering area of shallow bedrock; note ENE trending strike ridges of Hampden Basalt and East Berlin Formation.

5.5 Stop light at West Street. *If not going to Stop 8, turn right (east) on West Street and continue 1.2 miles to return to Dinosaur State Park.*
6.3 Turn left (west) on New Britain Ave., Route 160.
6.7 Turn right (north) on Hayes Road.
7.7 Turn left (west) on Two Rod Highway.
8.7 Cross Rocky Hill-Newington town line.
9.3 Stop light at Berlin Turnpike. Continue straight across turnpike onto Main Street, Route 176. We are now traveling over Newington ice-marginal deltaic deposits of glacial Lake Middletown.
9.9 Turn left (west) on New Britain Avenue, Route 174.
10.4 Turn left (south) on Willard Avenue.
10.6 Turn right (west) to continue on New Britain Avenue, Route 174.
11.1 Leave ice-marginal deposits and begin traveling through thin till overlying shallow bedrock-- NE trending strike ridges of Portland Arkose.
11.6 Descend into New Britain Spillway for glacial Lake Hitchcock.
11.8 Turn right (north) on Stamm Road. Park where we can.

Stop 8 . New Britain spillway for glacial Lake Hitchcock.

Water spilling from glacial Lake Hitchcock flowed through the New Britain spillway (Fig. 30) during the high levels of the lake (Connecticut phase 17.9 ka until 17.3 ka) and during its longer-lived stable phase (from 17.3 ka until 15.6 ka). This pathway was established during the time that the level of glacial Lake Middletown slowly lowered and exposed the lowest point across the drainage divide in the area. The deltaic deposits of Lake Middletown to the east in Cromwell and Rocky Hill (the Rocky Hill dam) are more extensive than those in New Britain and Newington, and hence this was the lowest way out. The spillway is approximately 2 miles in length and 800 ft in width. It is a swampy, underfit channel not occupied by any major stream. It was cut into Lake Middletown deposits, till, and incompetent shaley bedrock from an altitude of about 115 ft down to its present 58-ft level. A 24-ft water column in the channel produced the 82-ft stable level.

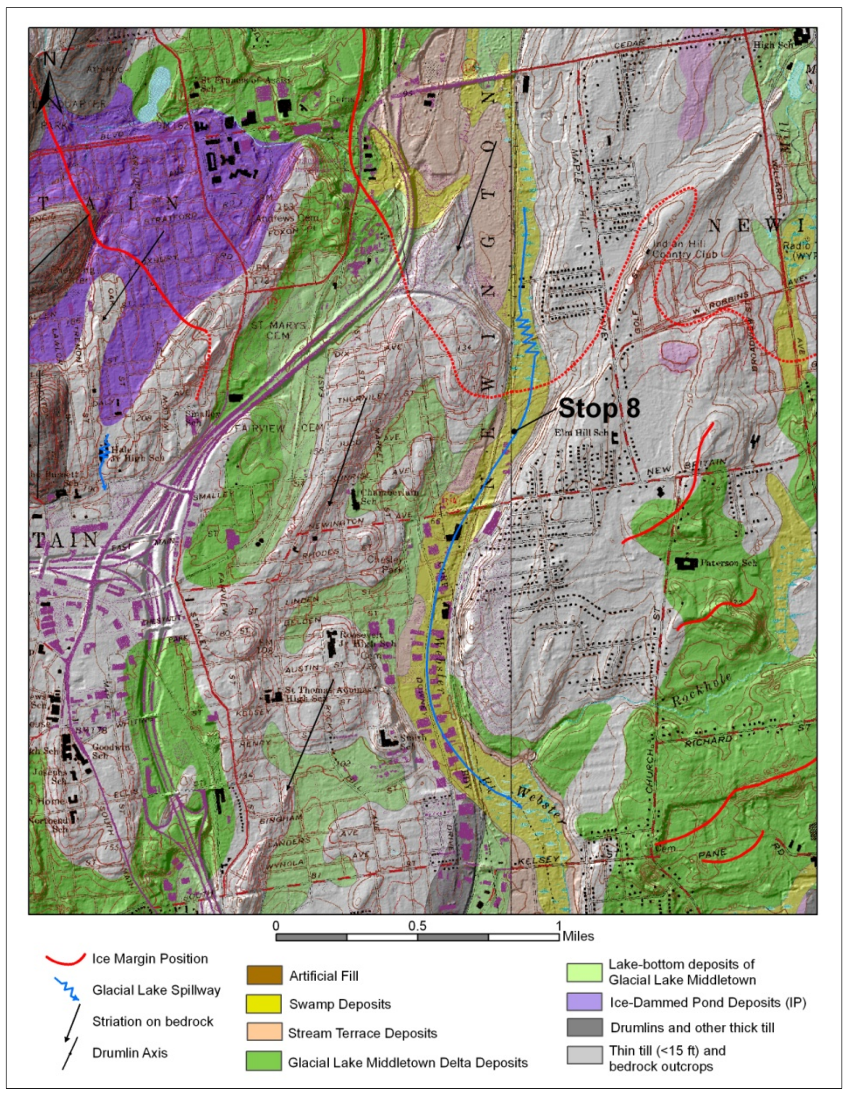


Figure 30. Surficial geology in the vicinity of the New Britain Spillway, Stop 8 (Stone and others, 2005).

Previous studies of glacial Lake Hitchcock have considered the New Britain spillway as the only base-level control for the lake. But lowering of the channel could only happen as base levels to the south were lowered. When glacio-isostatic depression is taken into account (Fig. 9), by the time that the spillway was at its stable level 82-ft level, there was no fluvial gradient south of the spillway, and sea level in Long Island Sound was rising. Therefore, Lake Hitchcock simply could not lower further until glacial rebound began.

The New Britain Spillway was abandoned at about 15.6 ka when the Rocky Hill dam was breeched. The lake lowered by about 20 ft and then quickly emptied south of the Holyoke Range in Massachusetts and Connecticut.

Mileage **Return to Assembly Point at Dinosaur State Park**

0.0 Return to New Britain Avenue, Route 174. Turn left, east.
1.1 Turn left (north) on Willard Avenue.
1.2 Turn right (east) to continue on New Britain Ave.
1.8 Turn right (south) on Main Street.
2.5 Continue straight across Berlin Turnpike onto Griswold Ave./Two Rod Highway.
4.0 Turn right (south) on Hayes Road.
5.0 Turn left (east) on Route 160
5.4 Turn right (south) on Route 3.
6.2 Turn left (east) on West Street
7.4 Turn right into Dinosaur State Park parking lot.

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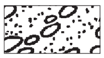
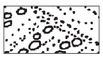
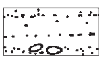
Explanation of Symbols for Appendix 1

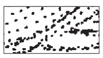
 Boulder gravel, planar bedded, massive, contains imbricate clasts

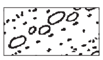
 Pebble-cobble gravel, planar bedded, massive

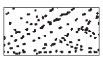
 Pebble-cobble gravel and sand, planar-tabular crossbedded

 Pebble gravel and sand, trough crossbedded

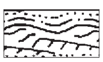
 Coarse sand, planar bedded

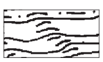
 Coarse sand, trough crossbedded

 Pebble-cobble gravel and sand in planar foreset beds

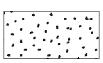
 Coarse sand in trough foreset beds

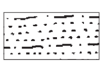
 Coarse sand in trough crossbeds

 Fine-medium sand in type A ripple cross laminations, which contain dipping lee-side laminations

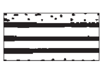
 Fine-medium sand in type B ripple cross laminations, which contain dipping stoss-side and lee-side laminations

 Fine sand and silt in draped laminations

 Fine-medium sand, massive

 Silt and very fine sand in parallel laminations

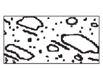
 Clay laminations

 Fine sand and silt in normally graded lamination

 Fine sand in solitary ripple cross laminations

 Dropstone

 Dropclot of nonsorted material

 Till

Reprinted from Quaternary Geologic Map of Connecticut and Long Island Sound Basin
Stone and others, 2005, U.S. Geological Survey, SIM 2784

Appendix 1. Sedimentary Facies and Morphosequences of Glacial Meltwater Deposits in Connecticut

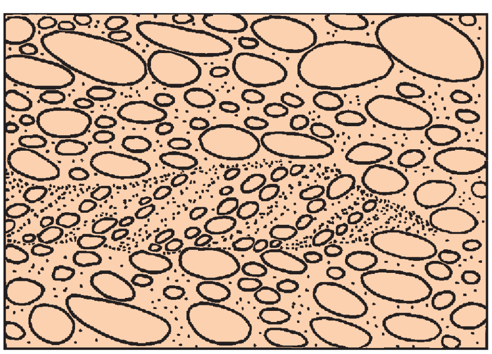
By Byron D. Stone and Janet Radway Stone

Sedimentary Facies

Sedimentary facies are defined on the basis of lithic characteristics of texture and sedimentary structure. In Connecticut, they are related to specific environments of deposition along the path of meltwater flow. Glaciofluvial sediments were deposited in meltwater streams, glaciodeltaic sediments were deposited where meltwater streams entered glacial lakes, and glacial lake-bottom sediments were deposited in deeper parts of glacial lake basins. For the illustrations that follow, please see the Explanation of Symbols on the facing page.

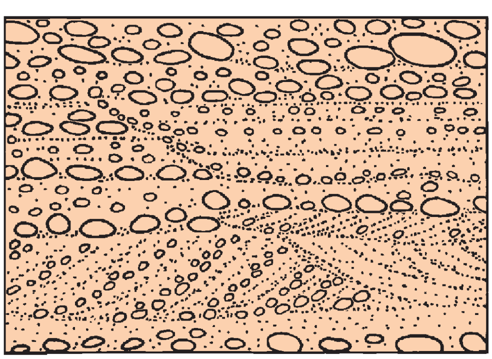
Glaciofluvial Sedimentary Facies

Sediments in glaciofluvial sedimentary facies are horizontally stratified alternating beds of gravel and sand that are either planar, or crossbedded. Gravel clasts are well rounded to subrounded and commonly are imbricated. The sand matrix in gravel beds is poorly sorted, medium to coarse sand. Gravel planar beds and sets of crossbeds are less than 5 ft thick; all beds are bounded by erosional contacts; individual beds extend laterally as much as 50 ft. The total thickness of sediments is commonly 15 ft in delta topset beds, and rarely as much as 100 ft in ice-marginal fluvial heads of deposits. Glaciofluvial sediments were deposited by meltwater streams at or in front of the ice margin as outwash in valleys with steeper gradients, in promorainal outwash plains, and in inset valley terraces distal from the ice margin. Ice-channel sediments were deposited by meltwater in ice-walled channels and tunnels. Glaciofluvial sediments are present in delta-surface plains as delta topset beds; these are included as part of the glaciodeltaic environment.



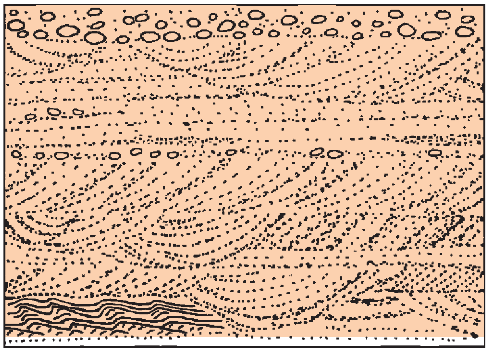
COARSE GRAVEL FLUVIAL FACIES

Cobble-boulder gravel beds, coarse sand matrix, massive planar bedded, planar-tabular crossbedded
Minor interbedded medium-coarse sand beds, planar-tabular and trough crossbedded, ripple laminated
No silt or clay beds
5 to 50 ft total thickness
At ice-marginal head of morphosequences



SAND AND GRAVEL FLUVIAL FACIES

Pebble-cobble gravel beds, massive planar bedded, planar-tabular and trough crossbedded
Interbedded medium-coarse sand beds, planar-tabular and trough crossbedded, ripple laminated
Minor sets of silt laminae, minor clay
3 to 50 ft total thickness
In outwash deposits and as delta topset beds



COARSE PEBBLY SAND FLUVIAL FACIES

Coarse pebbly sand beds, massive planar bedded, planar-tabular and trough crossbedded
Interbedded pebble gravel beds, and medium-coarse sand beds, planar bedded and ripple laminated
Minor sets of silt laminae, minor clay
0.5 to 25 ft total thickness
In distal outwash or delta topset deposits

SAND AND GRAVEL ICE-CHANNEL FLUVIAL FACIES

Pebble-cobble gravel beds, massive planar bedded, planar-tabular crossbedded
Interbedded medium-coarse sand, planar bedded, planar-tabular and trough crossbedded, ripple laminated
Minor sets of silt and clay laminae
10 to 100 ft total thickness
In ice-contact esker and ice-channel deposits

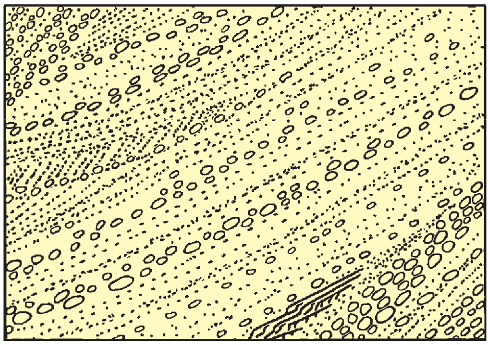
Glaciodeltaic Sedimentary Facies

Sediments in deltaic sedimentary facies include topset, foreset, and bottomset beds. Horizontally layered delta topset beds are glaciofluvial beds of sand and gravel deposited in the subaerial plain of the delta. These disconformably overlie steeply dipping delta foreset beds of sand and gravel that were deposited subaqueously. Subhorizontal delta bottomset beds of fine sand and silt, deposited subaqueously, intertongue updip with delta foreset beds. Glaciodeltaic sediments were deposited where meltwater streams entered glacial lakes directly from the ice margin or at the mouths of lake tributary streams.

Delta topset beds are flat-lying fluvial beds composed mainly of sand and gravel fluvial facies and coarse pebbly sand fluvial facies (described above as glaciofluvial sediments). Topset beds are commonly 10 to 20 ft thick in proximal sections and 0.5 to 3 ft thick at the distal edge of the delta plain. The altitude of the basal contact of topset beds is a close approximation to the surface altitude of the glacial lake.

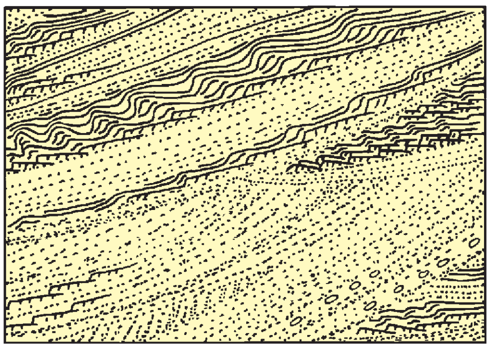
Delta foreset beds are steeply dipping (15°-30°) and are grouped in disconformable sets, commonly 2 to 20 ft thick, which may extend tens of feet down dip. Sets of foreset beds are planar tabular in proximal parts of deltas and trough-shaped in distal parts. Foreset strata contain planar, crossbedded, and ripple-laminated beds. Total thickness of foreset beds varies from 6 to 100 ft.

Delta bottomset beds are subhorizontally layered and contain alternating sets of ripple and planar laminations composed of medium to fine sand, silt, and clay. Bottomset beds are grouped in conformable flat-lying sets or in disconformable trough-fill sets that dip less than 10°. Total thickness of bottomset beds varies from 6 to 30 ft. Bottomset beds commonly extend gradationally from the lobate depositional front of the delta to the lake-bottom deposits.



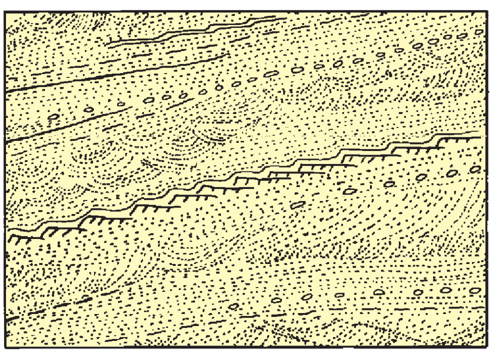
SAND AND GRAVEL DELTAIC FORESET FACIES

Pebble-cobble gravel in planar-tabular and trough foreset beds, parallel bedded, minor openwork gravel
Interbedded fine-coarse sand in trough foreset beds, parallel bedded, planar-tabular crossbedded, and ripple laminated
Some sets of silt laminae, minor clay, minor flowtills
6 to 100 ft total thickness
In proximal parts of deltaic deposits



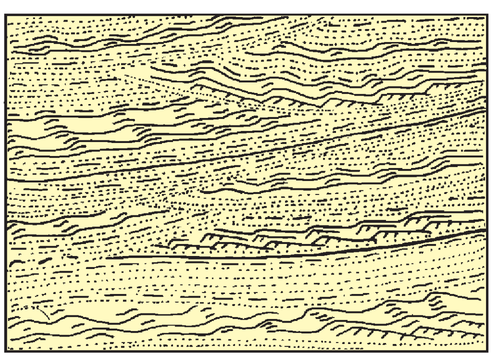
SANDY DELTAIC FORESET FACIES

Fine-coarse sand in planar-tabular and trough foreset beds, parallel bedded, ripple laminated
Interbedded fine pebble gravel in planar-tabular foreset beds, parallel bedded
Interbedded sets of draped silt and minor clay laminae
6 to 100 ft total thickness
In central and distal parts of deltaic deposits



SAND AND GRAVEL DELTAIC BOTTOMSET FACIES

Coarse pebbly sand and pebble gravel in trough bottomset beds, parallel bedded, trough and planar-tabular crossbedded
Interbedded fine-medium sand, parallel bedded, ripple laminated
Interbedded sets of draped silt and minor clay laminae
6 to 20 ft total thickness
In proximal and central parts of deltaic deposits

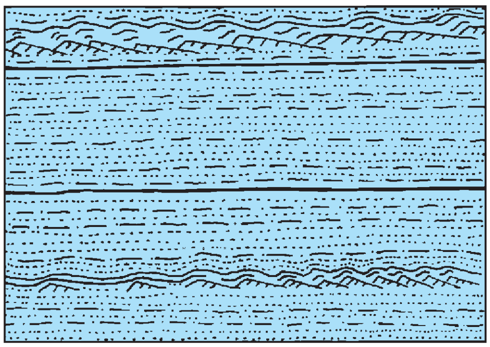


FINE SAND DELTAIC BOTTOMSET FACIES

Fine-medium sand in planar and trough bottomset beds, parallel bedded, ripple laminated
Interbedded sets of draped silt and minor clay laminae
6 to 30 ft total thickness
In distal parts of deltaic deposits

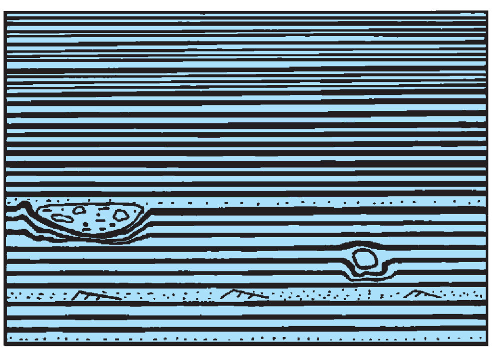
Glacial Lake-Bottom Sedimentary Facies

Sediments in glacial lake-bottom sedimentary facies include horizontally layered, laterally extensive, fine to very fine sand, silt, and clay deposited subaqueously on lake-bottom areas. Sandy lake-bottom sediments intertongue with deltaic or lacustrine fan bottomset beds and gradationally merge into silt-clay facies in large lake basins. Irregularly spaced laminations of fine sand, silt, and generally thin clay (0.2 in thick) merge into regularly spaced couplets of silt and clay (varves) in which silt layers are commonly 0.2 to 1.5 in thick and clay layers are 0.2 to 0.8 in thick. Lake-bottom sediments also include gently to steeply dipping gravel and sand beds in proximal parts of ice-marginal lacustrine fans, locally intertonguing with ice-tunnel deposits, and laterally extensive sand and silt beds in distal parts of lacustrine fans deposited on the bottom of deep lakes.



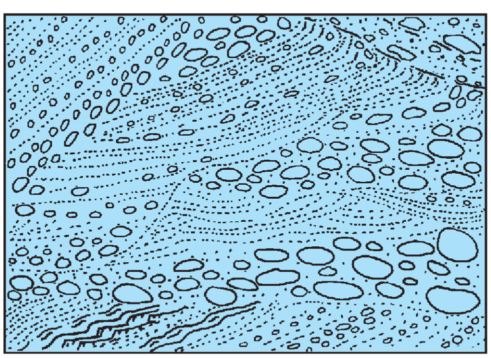
SANDY LAKE-BOTTOM FACIES

Fine sand-silt, irregularly spaced parallel laminae, ripple laminated
Interbedded sets of silt and clay laminae
10 to 60 ft total thickness
In lake-bottom deposits proximal to deltaic deposits



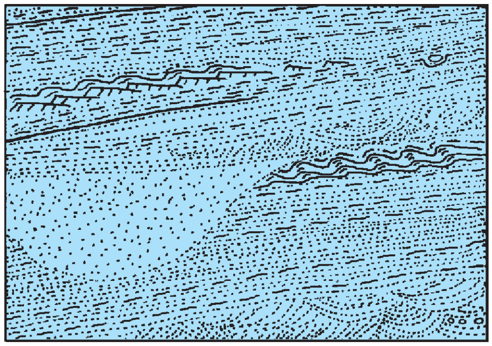
SILT-CLAY LAKE-BOTTOM FACIES

Silt-fine sand and clay in irregularly spaced parallel laminae or regularly spaced varve couplets, minor ripple laminated
Minor interbedded fine sand, parallel laminated, ripple laminated
6 to 200 ft total thickness
In distal lake-bottom deposits



SAND AND GRAVEL LACUSTRINE FAN FACIES

Pebble-cobble gravel, coarse sand, and minor flowtill in planar-tabular and trough foreset beds, parallel bedded, planar-tabular crossbedded
Local compact till at top of section
Minor interbedded fine-medium sand, parallel bedded, ripple laminated
Minor interbedded sets of draped silt and minor clay laminae
6 to 60 ft total thickness
In proximal parts of lacustrine fan deposits



SAND-SILT LACUSTRINE FAN FACIES

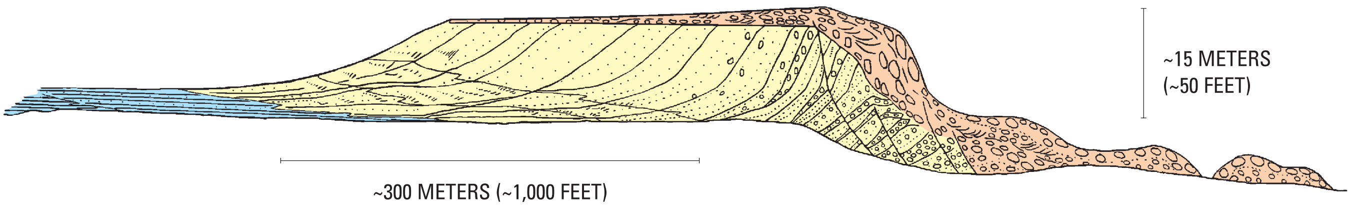
Fine-medium sand in planar and trough bottomset beds, parallel bedded, ripple laminated
Interbedded sets of draped silt and minor clay laminae
6 to 60 ft total thickness
In distal parts of lacustrine fan deposits

Morphosequences

Morphosequences are bodies of stratified meltwater sediments that are contained in a related series of landforms and that are mappable as individual deposits at detailed scale (maps at 1:24,000 scale; Jahns, 1941; Koteff and Pessl, 1981). Each morphosequence is an assemblage of sedimentary facies that were deposited contemporaneously. Each morphosequence is a small lithologic unit that was built in close association with the ice margin in a particular section of a valley as meltwater streams aggraded their beds and (or) supplied deltaic and lacustrine sediment to proglacial lakes and ponds. The surface altitude of fluvial sediments in each morphosequence was controlled by a specific base level, commonly a glacial lake water plane. Morphosequences each consist of a proximal part (head) deposited within or near the ice margin, and a distal part deposited farther away from the ice margin. Both grain size and collapse deformation of beds decreases from the proximal to the distal part of each morphosequence. Ice-marginal morphosequences were deposited in contact with the ice margin. The heads of many ice-marginal morphosequences extended well up into the ice margin in channels and tunnels; melting of adjacent and subjacent ice caused the collapse of these headward sediments to lower positions where they commonly were buried by later sediments. The ice-margin position lines (shown on the map for many ice-marginal morphosequences) represent the scarp between the severely collapsed headmost part of the deposit and the part that retains some of the flattish top at or close to the original level of deposition. That ice-margin collapse scarp defines the outermost fringe of continuous ice at the time that deposit was built. Near-ice-marginal morphosequences were deposited short distances in front of the ice margin, separated from it by valley segments too steep for deposition. Few morphosequences extend distally more than 5 mi, and most are less than 1 mi in length.

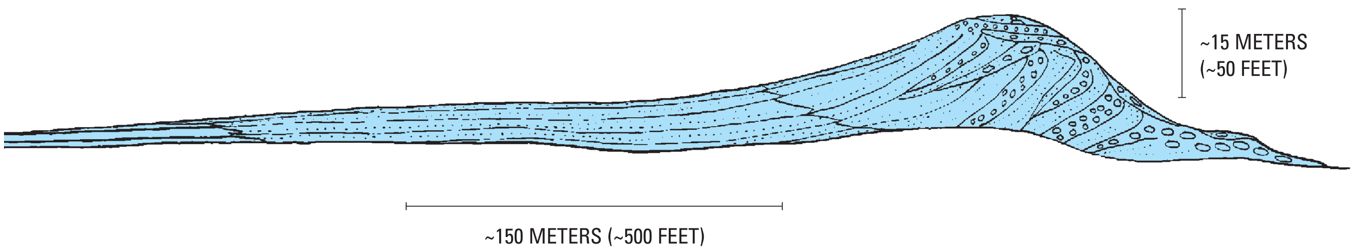
In any one valley, individual morphosequences were deposited sequentially as the ice margin retreated systematically northward. Consequently, in many places the distal, finer grained facies of a younger morphosequence stratigraphically overlies the proximal, coarse-grained facies of a preceding morphosequence.

Seven types of morphosequences are recognized in Connecticut, the names of which have been modified from the original terminology of Koteff and Pessl (1981). The names refer to the depositional environment of each morphosequence type. The head of each morphosequence is either ice marginal (ice contact) or near ice marginal. Morphosequences consist of combinations of glaciofluvial, glaciodeltaic, and glacial lake-bottom sedimentary facies; they can occur in various depositional systems: major ice-dammed lakes (IL), major sediment-dammed lakes (SL), related series of ice-dammed ponds (IP), related series of sediment-dammed ponds (SP), proximal meltwater streams (FP), and distal meltwater streams (FD).



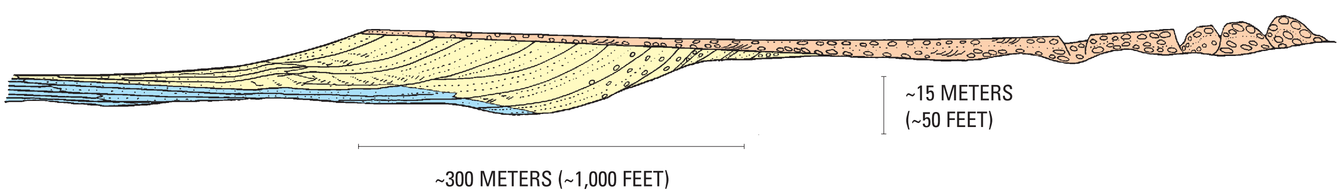
Ice-Marginal Deltaic Morphosequence

This type of morphosequence consists of a delta deposited in contact with the ice margin. Collapsed coarse-grained topset and foreset beds are present in the proximal part; noncollapsed finer grained topset, foreset, and bottomset beds are present in the distal part. Ice-marginal deltas that were built into larger glacial lakes (lakes not completely filled by deltaic sediments) preserve depositional frontal slopes that merge with lake-bottom sediments beyond the lobate front. Esker-fed deltas are ice-marginal deltas that have coarse, fluvial ice-channel sediments preserved in esker form within the ice-marginal zone. This type of morphosequence occurs in the IL, SL, IP, and SP depositional systems.



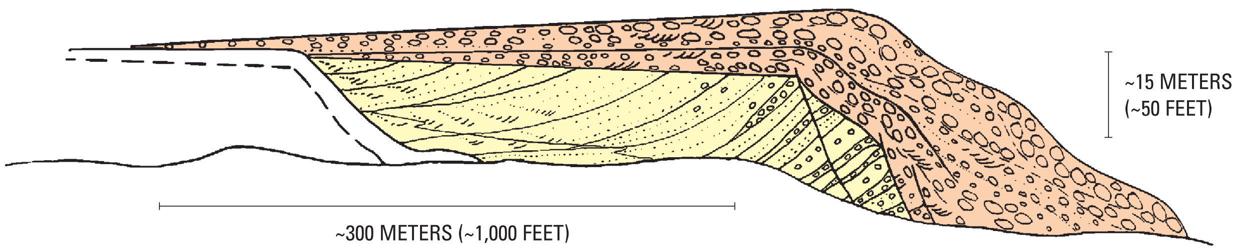
Ice-Marginal Lacustrine Fan Morphosequence

This type of morphosequence consists of coarse-grained lacustrine fan beds deposited in contact with the ice margin on the bottom of a glacial lake. Coarse-grained sediments, interbedded with or overlain by compact till and probably glaciotectonically deformed, are present in the proximal part. Finer grained beds are laterally extensive in the distal part; in some places, these beds grade distally into a varved lake-bottom facies. Lacustrine fans prograded onto the bottom of deep glacial lakes from the point of contact between the ice margin and the lake bottom at the mouths of subglacial tunnels. They are common in central Long Island Sound, in the SL depositional system.



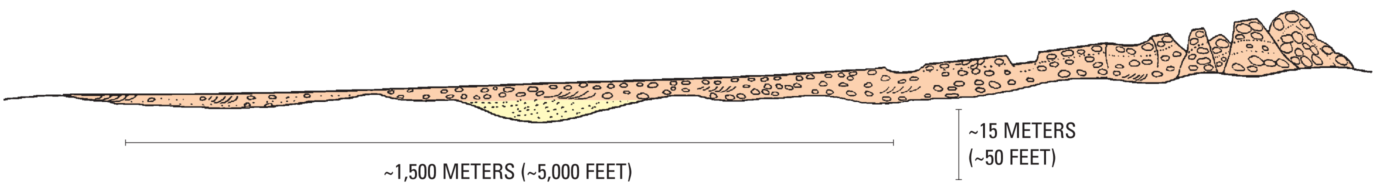
Ice-Marginal Fluviodeltaic Morphosequence

This type of morphosequence consists of fluvial sediments deposited in contact with the ice margin that stood upslope from a glacial lake or pond. Collapsed, coarse-grained fluvial beds occur proximally; noncollapsed deltaic topset, foreset, and bottomset beds occur distally. If the glacial lake or pond was not completely filled by a delta, lake-bottom sediment lies in front of as well as beneath the frontal slope of the delta. This type of morphosequence occurs in the IL, SL, and rarely in the SP depositional systems.



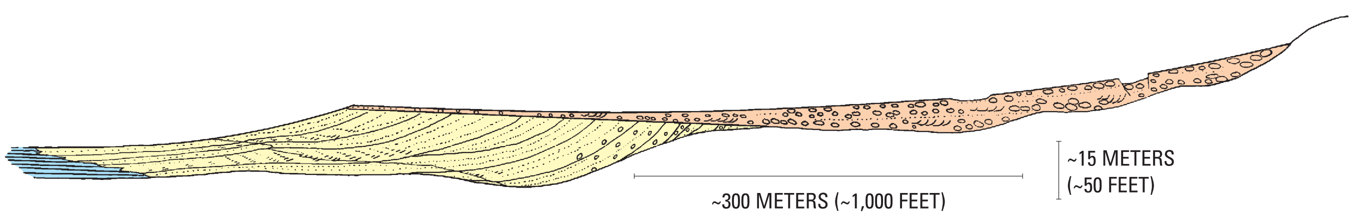
Ice-Marginal Deltaic-Fluvial Morphosequence

This type of morphosequence consists of deltaic and fluvial beds deposited in contact with the ice margin. Collapsed, coarse-grained beds occur proximally; noncollapsed beds occur distally. In small glacial lakes and ponds that were completely filled with deltaic beds, continued aggradation produced thick fluvial sediments which completely cover delta topset beds. The fluvial beds in some places actually extend distally through the lake spillway and overlap onto older deposits to the south. This type of morphosequence occurs principally in the SP and rarely in the IP depositional systems.



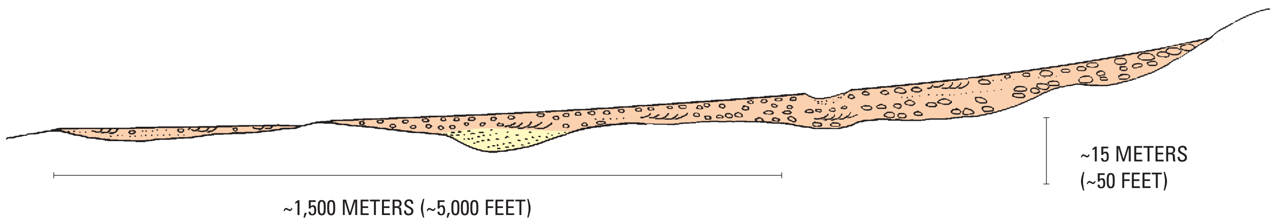
Ice-Marginal Fluvial Morphosequence

This type of morphosequence consists of fluvial beds deposited in contact with the ice margin. Collapsed, coarse-grained fluvial sediments occur in the proximal part; noncollapsed, finer grained fluvial sediments occur distally. These morphosequences were deposited in valleys that were too steep for ponding of meltwater to occur; these deposits are relatively rare in Connecticut. They occur in the FP depositional system.



Near-Ice-Marginal Fluviodeltaic Morphosequence

This type of morphosequence consists of fluvial beds deposited in tributary valleys but not in direct contact with the ice margin, and deltaic sediments deposited downstream in a lake or pond. Coarse-grained fluvial beds are in tributary valleys and commonly are collapsed in segments related to the melting of local ice blocks. Deltaic sediments include sand and gravel topset beds and coarse- and fine-grained foreset beds. Deltaic sediments generally are not collapsed. This type of morphosequence is common in the SL and SP depositional systems, and is in some units in the IL system.



Near-Ice-Marginal Fluvial Morphosequence

This type of morphosequence consists of fluvial beds deposited in valleys but with heads not in direct contact with the ice margin. Coarse-grained glaciofluvial beds commonly are collapsed in segments related to the melting of local ice blocks. Distal parts of deposits contain gravel and sand-dominated facies, and are little collapsed. Some deposits include subsurface fine-grained sediments of small ponds in local closed basins and fine-grained kettle-fill sediments. This type of morphosequence is the principal component of the FD depositional system.

Other Deposits

Three other types of meltwater deposits are common in Connecticut, but these are not morphosequences because they were not deposited in close association with the ice margin. These deposits also are composed of various sedimentary facies.

Lake-Bottom Deposits

Fine-grained, lacustrine beds accumulated on the bottom of glacial lakes and ponds, locally in contact with the ice margin. Fine sand and silt sediments intertongue with deltaic or lacustrine fan bottomset beds; distally, the sand and silt beds grade into silt and clay varved sediments. Lake-bottom deposits are extensive distally from deltaic morphosequences within the IL and SL depositional systems. Fine-grained silt and clay settled out of suspension continuously during the life of these large lakes, and is therefore associated with the deposition of multiple deltaic morphosequences. In smaller lakes of the IP and SP depositional systems, lake-bottom sediments are the distal facies of deltaic morphosequences.

Meltwater Terrace Deposits

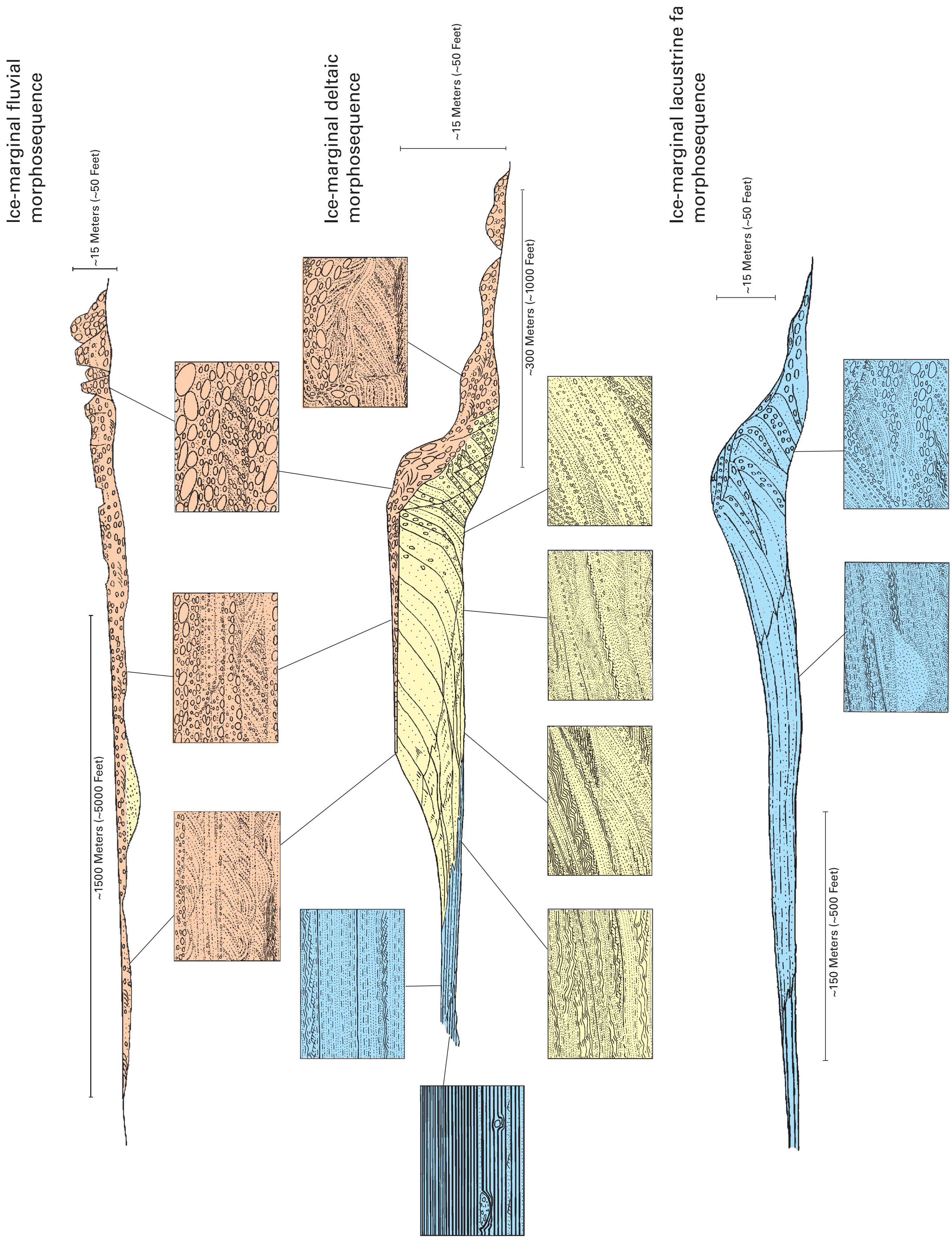
Meltwater terrace deposits consist of fluvial beds deposited as erosionally inset terraces, generally not traceable to an ice-marginal head. Sediments commonly are lithologically distinct from the older and higher deposits which are entrenched, and generally are not collapsed. This type of deposit occurs in the FD depositional system.

Meteoric Fluviodeltaic Deposits

This type of deposit consists of fluvial and deltaic beds deposited disconformably over older glaciodeltaic and glaciolacustrine sediments by meteoric streams in glacial Lake Hitchcock. Fluvial beds consist mainly of sand and grade to deltaic plains underlain by thin sand topset beds and thin sand delta foreset and bottomset beds. This type of deposit was emplaced only along courses of major streams that entered glacial Lake Hitchcock after deglaciation of tributary drainage basins in the SL depositional system.

Facing page. Diagrams showing sedimentary facies relationships between three types of morphosequences. The distribution of, thickness of, and relationships between stratigraphic sedimentary facies are shown schematically with similar symbols in each morphosequence.

Appendix 1


Ice-marginal fluvial morphosequence
~15 Meters (~50 Feet)
~1500 Meters (~5000 Feet)
Ice-marginal deltaic morphosequence
~15 Meters (~50 Feet)
~300 Meters (~1000 Feet)
Ice-marginal lacustrine fa morphosequence
~15 Meters (~50 Feet)
~150 Meters (~500 Feet)