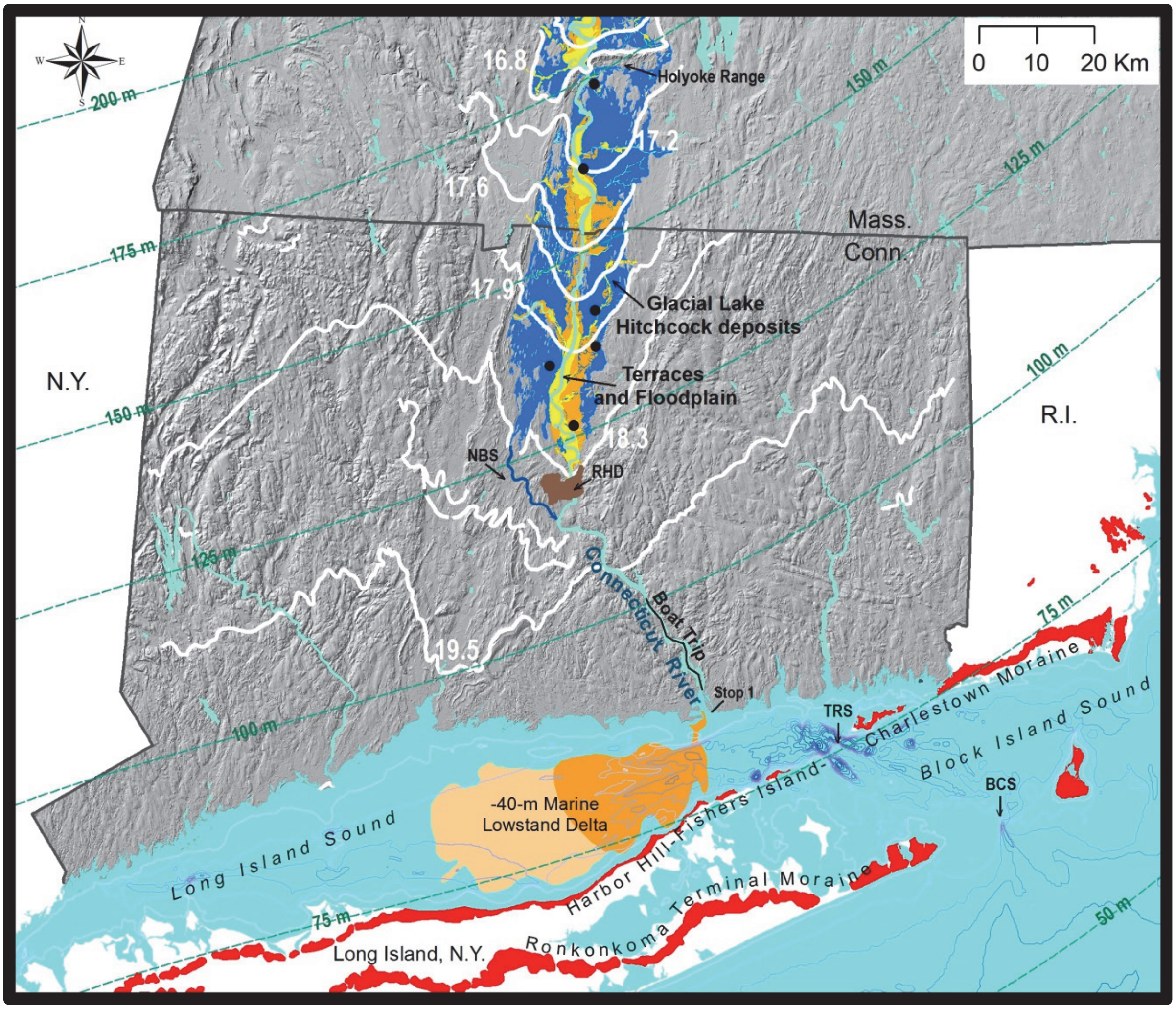
Glacial Lake Hitchcock and the Sea

Fieldtrip Guidebook for the 78th Annual Reunion of the Northeast Friends of the Pleistocene

Rocky Hill, Connecticut, June 5-7, 2015



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Cover: Glacial Lake Hitchcock in the Harford-Springfield Basin and the -40-m lowstand marine delta in Long Island Sound**.** **NBS**-New Britain spillway, **RHD**-Rocky Hill dam, **TRS**-The Race spillway, **BCS**-Block Channel spillway. Green lines are isobases of glacio-isostatic depression, numbers are total depression in meters. White lines are selected retreatal ice-margin positions, numbers are calibrated dates in thousands of years based on varve records from the North American Varve Chronology (Ridge and others, 2012). Black dots are locations of varve cores that penetrate to the base of the lake section and identify the oldest varve present at each site.

Northeast Friends of the Pleistocene

78th Annual Fieldtrip

June 5-7, 2015

Glacial Lake Hitchcock and the Sea

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Glacial Lake Hitchcock & the Sea

Friends of the Pleistocene Fieldtrip Conference, June 5-7, 2015

The Friends of the Pleistocene gathering is being held in Connecticut for the first time since 1935 when Richard Foster Flint hosted the 2nd annual FOP fieldtrip at New Haven, and stops included the Hartford Clay of Glacial Lake Hitchcock (at that time unnamed). Eighty years later, there is more to tell about Lake Hitchcock in Connecticut, thanks to geologists like Ernst Antevs, who in the 1920's gave us the powerful chronologic tool of varve correlation; and Richard Lougee, who understood early on why this lake existed, and in 1935 named it Glacial Lake Hitchcock. In recent years the regional compilation of many detailed on-land mapping studies (Hartshorn and Koteff, 1967; Koteff and Larsen, 1989; Jahns, 1966) combined with high-resolution offshore mapping (Stone and others, 2005), and recent calibration of the North American Varve Chronology (Ridge and others, 2012) have provided many new insights.

The fieldtrip will demonstrate the evidence for a close connection of Lake Hitchcock levels with lake levels and the position of sea level in Long Island Sound via a channel cut into glacial lake deposits in the lower Connecticut River valley, which is superposed on a bedrock ridge at the mouth of the Connecticut River. On the trip we will explain important offshore features like an extensive -40-m marine delta, and the altitudes of "The Race" spillway cut through the Harbor Hill moraine, Block Channel spillway cut through the terminal moraine, and the -85-m Block Delta built into Last Glacial Maximum (LGM) eustatic sea level 115 km south of the terminal moraine. The history of lake levels and knowledge of eustatic sea levels provided by the Barbados sea level curve (Bard and others, 1990) have implications for the magnitude of glacio-isostatic depression and the timing of rebound.

We will also review recent refinements to the chronology of ice retreat through the region as a result of new varve cores and the newly calibrated North American Varve Chronology (NAVC) (Ridge, 2004, Ridge and others, 2012) and discuss implications for the timing and mechanism of glacial Lake Hitchcock drainage in Connecticut.

Itinerary

Day 1

9:30 AM: Meet at Dinosaur State Park. Park cars, board bus.

10:30-11:30: Stop 1, Griswold Point, Old Lyme

11:45-12:15: Stop 2, DEEP Marine Headquarters at I-95 bridge, Old Lyme

12:15-1:45: Board RiverQuest Cruise Boat for tour up CT River to East Haddam, Stop 3 (Lunch included).

1:45-5:00: Stops 4-6, Drive north along CT River with Stops at Haddam Meadows, River Road to view "the Straits" of the Connecticut River, and into Middletown basin for Stop 6 on lake-bottom surface with rimmed depressions. Return to Rocky Hill.

6:00 PM: Happy Hour gathering on the patio at Al Garve restaurant with trip overview posters.

Day 2

8:00 AM: Meet at Dinosaur State Park. Board bus.

8:30-4:30: Fieldtrip stops 7-12 at view of Rocky Hill Dam for glacial Lake Hitchcock, New Britain Spillway, several stops in the Farmington delta complex, and Matianuck Dunes State Preserve for lake-bottom surface-- varves, dunes, and lithalsa scars. Lunch included.

6:00: Happy Hour and Buffet Dinner at Dinosaur State Park

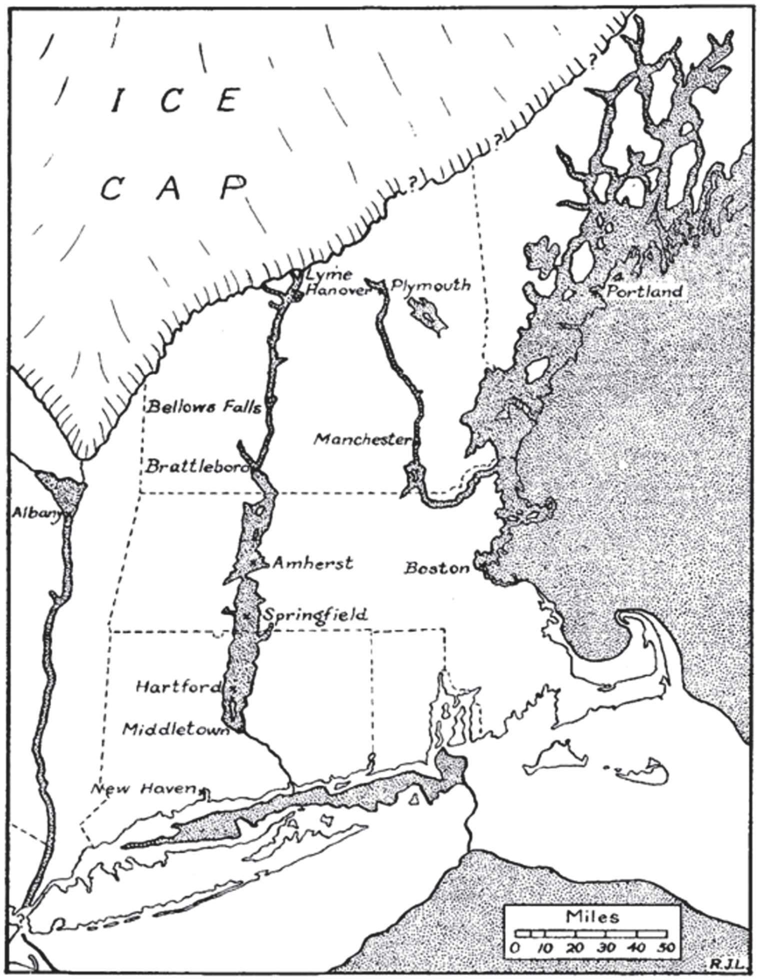
Day 3

8:00 AM: Assemble cars

8:30: Stop 13-- Glastonbury Bulky Waste Disposal excavation into Rocky Hill dam deposits.

10:00: Stop 14—Riverfront Park in Glastonbury to view varve cores.

11:30: Stop 15-- Redlands Brick Co. (formerly Kelsey-Ferguson) clay pit.



Map showing late-glacial conditions in New England when Lake Hitchcock was at its maximum extent in the Connecticut Valley. Lakes occupied the Hudson and Merrimac Valleys and Long Island Sound; the continental shelf of southern New England was dry land; and the sea covered much of the coastal belt north of Boston.

"...Lake Hitchcock owed its existence to two causes. First, the valley in the vicinity of the narrow gorge of the river at Middletown, Connecticut, was obstructed by a natural dam of glacial deposits which alone would have produced a lake. Second, the weight of the ice pressed down the crust of the earth so that the floor of the valley actually sloped northward toward the ice, effectually preventing the free flow of a river. As a result the lake began to appear as the ice uncovered Middletown, and it expanded into Massachusetts and Vermont and New Hampshire with increasing depth as the ice withdrew. In Vermont and New Hampshire the lake was only two or three miles wide, but in the broad lowland of Massachusetts and Connecticut it attained a width of ten or twelve miles and contained several large islands that are now lofty hills. In this broad southern portion of the lake which was earliest uncovered and where the sweep of winds was greatest the waves cut deeply into the island hills. West of Hartford, Connecticut, the cliffs and boulder-strewn terraces carved by the waves are well developed along the ancient shore. North of Connecticut the best features making the shore line are the deltas built by tributary streams where they joined the lake. These deposits occur as terraces high on the sides of the valley and close to the streams that built them. Their flat tops mark approximately the water level, and excavations reveal their characteristic inner structure of horizontal gravel beds overlying steeply dipping beds of sand."

Richard J. Lougee "Hanover Submerged" 1935

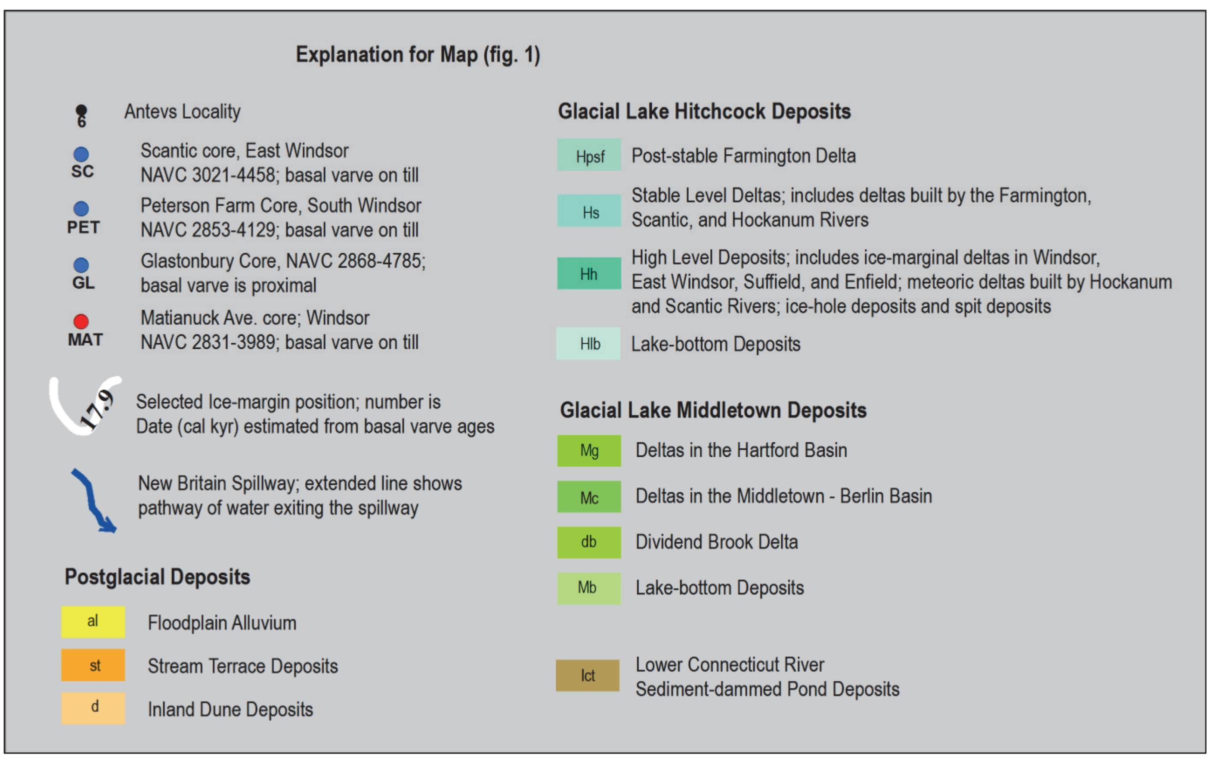
Glacial Lake Hitchcock in Connecticut and the New England Varve Chronology

The New England Varve Chronology

An approximate 4,000-year life span for glacial Lake Hitchcock was indicated by Ernst 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 assumes 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. Varved silts and clays of glacial Lake Hitchcock were used by Antevs (1922; 1928) to construct the New England varve chronology (NEVC). Antevs localities 1-6 (former clay pits), and 139 and 140 (stream bluffs) used to construct the chronology are in Connecticut. NE varve years 2868 to 3871, a total of 1003 years, are recorded at the eight Connecticut localities (fig. 1 and 2).

More recently, new varve records from long cores and many new radiocarbon dates have allowed for the correction, expansion, and calibration of Antevs' NEVC (Ridge, 2004; Ridge and others, 2012). The new formulation, called the North American Varve Chronology (NAVC) is a 5659-year varve sequence (AM years 2700-8358, which includes varves in the Hudson and Merrimack Valleys) that spans much of the last deglaciation, from 18,200 to 12,500 cal years BP.

In Connecticut, new long cores at four sites (figs. 1 and 2) at Glastonbury (GL), Peterson Farm (PET), Scantic (SC), and Matianuck (MAT) have allowed the expansion and correction of the varve sequence in the southern part of glacial Lake Hitchcock to include AM-2831 through AM-4785. NAVC age calibration of these varve years is 17,939 cal years BP for the first varve resting on till at the Matianuck core site (now the oldest observed varve from Lake Hitchcock), and 15,985 cal years BP for the uppermost varve in the Glastonbury core– a total of 1954 years. Previously, the oldest varve recorded by Antevs (1928) in the southern end of Lake Hitchcock was AM-2868 (Salmon Brook in Glastonbury, site 139), by coincidence also the bottom varve in the new Glastonbury core.



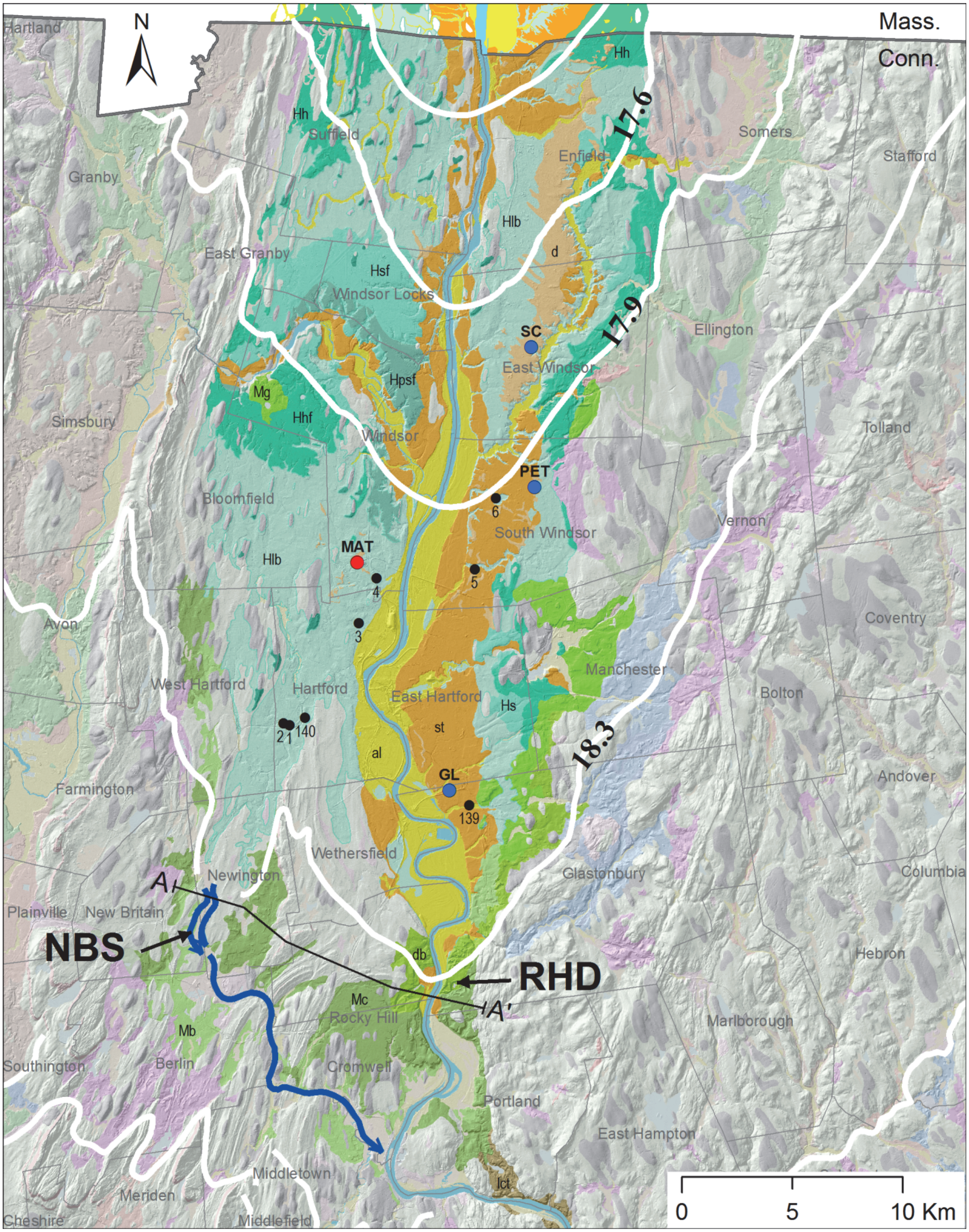


Figure 1. Map showing glacial Lake Hitchcock deposits and glacial Lake Middletown deposits in Connecticut, modified from Stone and others (2005). Line A-A' is section line for figure 3. **NBS**–New Britain Spillway; **RHD**–Rocky Hill Dam. White lines are retreatal ice-margin positions; numbers are estimated dates in thousands of years.

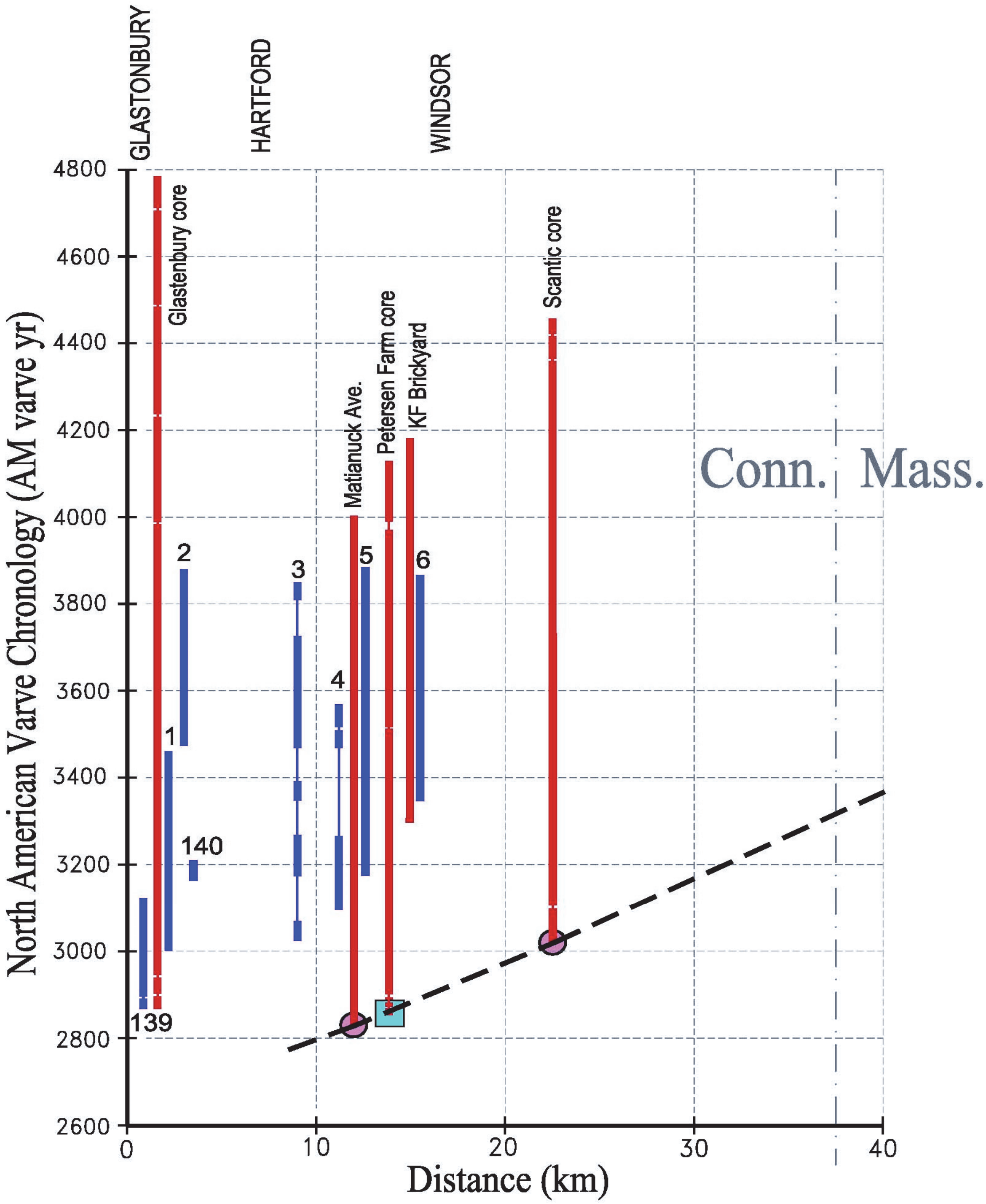


Figure 2. Composite cross section of the Connecticut Valley from Glastonbury in the south to the Massachusetts border showing all measured varve sections in the southern basin of Lake Hitchcock (locations on fig. 1) and their relative ages in NAVC varve years. Antevs' (1922, 1928) original sections are shown in blue and more recent measurements shown in red are from drill cores except the KF Brickyard. The pink circles represent sections where a basal varve (bottom varve in the sequence resting on till) has been recorded while the blue square represents ice-proximal varves. The dashed line represents the inferred time of ice recession along the axis of the valley at a rate of ~55 m/yr.

Glacial Lake Hitchcock

Glacial Lake Hitchcock (fig. 1) occupied the Hartford Basin in Connecticut and southern Massachusetts for about 2500 years, beginning at about 18 cal ka and continuing until about 15.5 cal ka. Deposits of this lake include ice-marginal deltas, meteoric-water-fed deltas, and extensive lake-bottom sediment consisting of varved silt and clay. The lake was impounded behind a mass of previously emplaced deltaic deposits that blocked the CT river valley between Rocky Hill and Glastonbury. These deltas were built into a slightly earlier lake, glacial Lake Middletown (fig. 1). This mass of meltwater deposits is often referred to as the "Rocky Hill dam." The dam is largely in place today with surface altitudes at 46-49 m, except where the modern Connecticut River cuts through (fig. 3). The spillway for glacial Lake Hitchcock was not over the dam however, but was to the west along the New Britain-Newington town line at the lowest place across the Mattabesset River drainage divide which separates the Hartford and Middletown basins (figs. 1 and 3). The 20-m deep, 250-m wide, 6.5-km long, under-fit channel contains no through-going stream today; most of the spillway is underlain by a swampy drainage divide, and the southern part contains Webster Brook, a small south-flowing tributary to the Mattabesset River (see stop 8).

When the ice margin first retreated north of that divide into the Hartford basin, glacial Lake Middletown water covered the later New Britain spillway location. Early ice-marginal deltas in the Hartford basin were controlled by glacial Lake Middletown. When Lake Middletown dropped to below 35 m, the New Britain spillway area emerged and became the spillway for glacial Lake Hitchcock as a separate water body. This happened at about 17.9 cal ka, when the ice margin was at Windsor and East Windsor (fig. 1). During the early life of glacial Lake Hitchcock, the New Britain spillway (NBS) was eroded into till and older stratified deposits so that water levels at the spillway lowered from about 35 m down to 25 m in altitude (Langer, 1977; Langer and London, 1979). Since the bedrock floor is at 18 m altitude today, water in the channel was about 7 m deep when the spillway level reached 25 m in altitude.

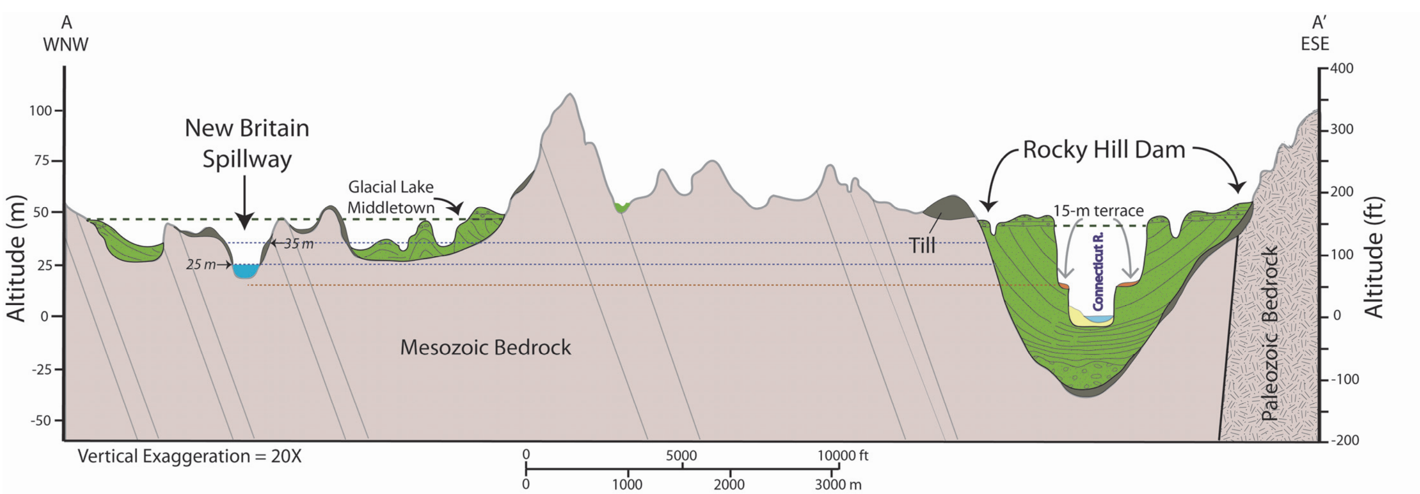


Figure 3. West to East Cross section between the New Britain spillway and the Rocky Hill dam. See line of section A-A' (fig. 1).

In Connecticut, all ice-marginal deltas record lake levels higher than the stable level. These deltas show a gradual lowering of lake level as the ice retreated northward and the New Britain spillway was incised down to bedrock (fig. 1). Ice-marginal deltas in Windsor and East Windsor record 35-m levels at the spillway. To the north, ice-marginal deltas in Suffield and Enfield indicate 33-m levels at the spillway; still farther north in Suffield and Enfield, the Shea Corner and Enfield deltaic deposits record levels just below 30 m at the New Britain spillway. This early phase of glacial Lake Hitchcock is recorded by ice-marginal deltas that extend into southern Massachusetts and were built to lake levels between 29 and 26 m projected to the spillway. This higher level phase of the lake has been referred to as the "Connecticut Phase" (Koteff and others, 1988).

Delta levels in Massachusetts indicate that a stable lake level, 25 m in altitude, had been reached by the time the ice margin had retreated to just north of the Chicopee River valley. The newly calibrated New England varve chronology (Ridge and others, 2012) places the ice front in this position at about 17.3 cal ka. Altitudes of topset-foreset contacts of ice-marginal deltas, from southern Massachusetts to near the Massachusetts–New Hampshire border, project to the stable 25-m level at the New Britain spillway on a straight line, which is tilted up to the north-northwest at a slope of 0.9 m/km. The linearity of these projected delta altitudes indicates that the lake level was stable during the time of ice retreat from Chicopee, MA to Turners Falls, MA, and was not affected by glacio-isostatic tilting until all of the deltas had been constructed. The ice margin was at Turners Falls at 15.6 cal ka (varve year NAVC-5200), when the lake level dropped by 5-6 m and this event is recorded by lowered inset levels in a large part of the Montague delta.

Deltas not associated with the ice margin, built by meteoric water from most river valleys that entered the lake also project to the stable lake level. In Connecticut, these include deltas at the Hockanum River, Scantic River, and where the Farmington River constructed a large delta northeastward into the lake in the area now surrounding Bradley International Airport.

The stable level of Lake Hitchcock ended when glacio-isostatic rebound began about 15.7 cal ka and the Rocky Hill dam was breeched. A preserved 15-m terrace (fig. 3) inset into the Rocky Hill dam sediments on both sides of the present Connecticut River in Rocky Hill and Glastonbury records a post-stable phase which was relatively brief in Connecticut. As rebound progressed, the dam was further entrenched from the 15-m level to about -5 m which is the base of floodplain alluvium today.

The Farmington River Delta Complex

The Farmington River delta complex (fig. 4) covers about 80 km^2 and records the entire history of lake levels in the Hartford-Springfield Basin. It consists of four deltaic deposits built into different lake levels. The first and highest deposit (**Mw**, on fig. 4) is ice-marginal and its level was actually controlled by glacial Lake Middletown. The deltaic deposit contains several large kettles and its front is collapsed; it may have been deposited in a large ice hole.

The high-level glacial Lake Hitchcock Farmington delta (**Hhf**, on fig. 4) covers about 25 km^2 and was deposited as the ice margin stood along its northeastern edge and prograded southwestward into the lake. The topset-foreset contact lies at 54 m; this level is equivalent to 35 m (the earliest high level) at the New Britain spillway when adjusted for glacio-isostatic tilting. The delta was built beginning at about 17.9 cal ka as recorded in the varve section at the Matianuck core site about 2 km south of the delta complex. The earliest varved sediment settled out in front of the prograding high-level delta; varve AM 2831 rests on till and is calibrated to 17,939 years (Stone and Ridge, 2009; Ridge and others, 2012).

The stable level Farmington delta (**Hsf**, fig. 4) covers about 50 km^2 and was constructed after the ice margin had retreated from the area. It built northeastward into the lake, fed by distal meltwater flowing from the Farmington valley to the west through the Tarrifville Gap. The topset-foreset contact is at 47 m, which records a 25-m level at the New Britain spillway. The extensive delta was built during the long stable phase of the lake from about 17.3 until 15.6 cal ka. The delta surface is tilted up to the N 21°W at 0.9 m/km, another indication that the lake was not affected by glacio-isostatic tilting until after its deltas in Connecticut and Massachusetts were constructed.

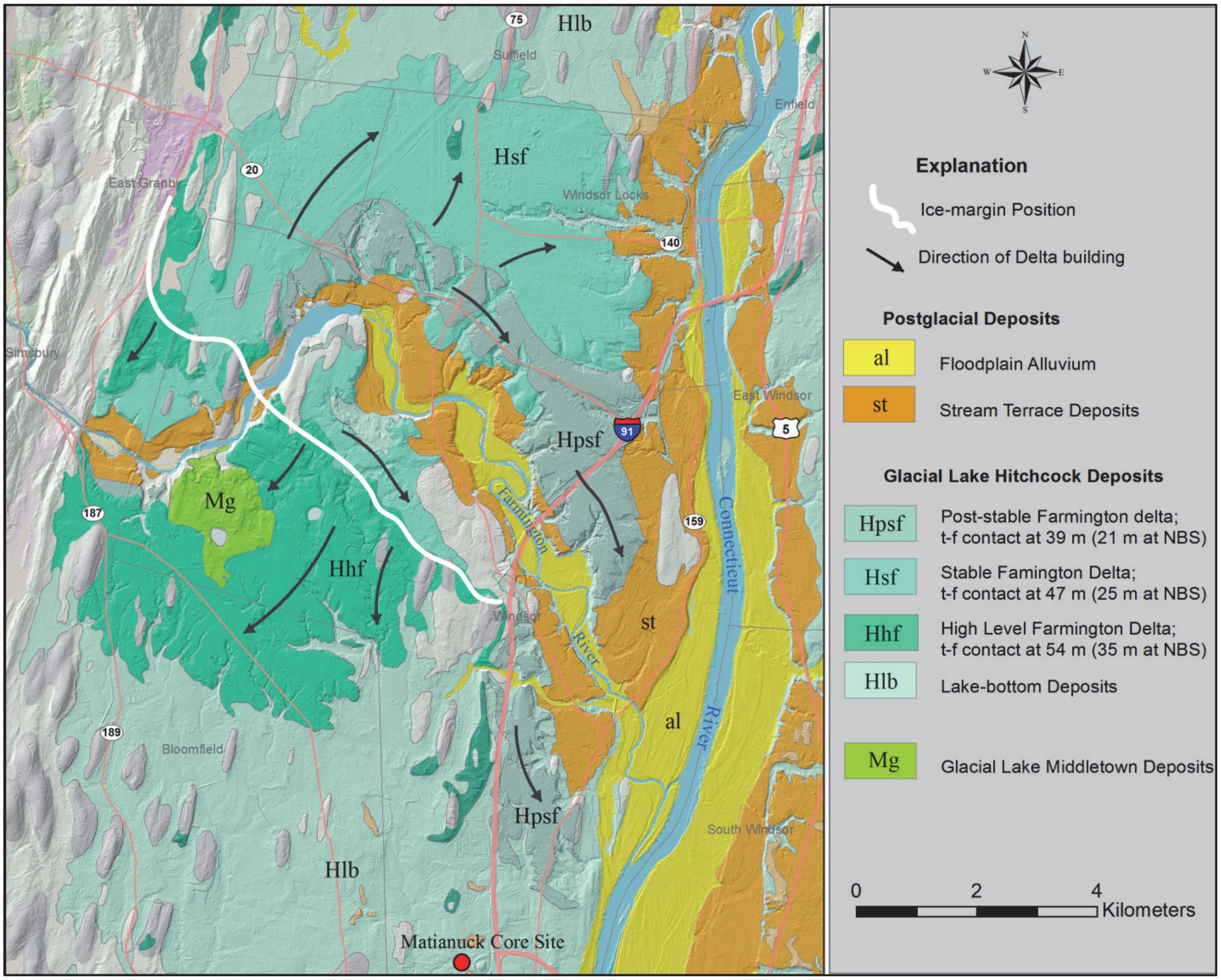


Figure 4. Map showing parts of the Farmington River delta complex.

Fluviodeltaic deposits (Hpsf) built southeastward into the lake by the Farmington River are less extensive. The paleo-Farmington River cut into the high level and stable level delta deposits and built a delta southeast ward into a lower lake level. A topset-foreset contact in the delta is at 39 m; this level projects to 21 m at the New Britain spillway, 4 m below the stable level. As we will see in later discussion, this was not possible until the dam broke and glacio-isostatic rebound began. The post-stable delta deposits record a “Post-stable Phase” (Koteff and others, 1988) of the lake during which levels were lower than the 25-m level of water spilling through the New Britain channel. A topset-foreset contact in the post-stable deltaic deposits north of the Farmington River is at 39 m; delta-surface altitudes farther south in the unit indicate slightly lower water levels. These levels project to levels of about 18-15 m at the Rocky Hill dam and record lowering of lake levels as the dam was entrenched.

Soon after deposition of the post-stable delta, the lakebed (surface of **Hlb**, fig. 1, 4) was exposed and contained a very high water table. The surface is extensively marked by circular to subcircular, rimmed depressions previously interpreted as open-system pingo scars (Stone and Ashley 1992). Today we believe a better interpretation of their genesis is that they are lithalsa remnants (Hugo and Geiss, 2007), which form typically in silty material on drained lake beds through the process of cryosuction in a discontinuous permafrost zone (Pissart, 2002; 2003; Calmels and others, 2008). In either case, the presence of these rimmed depressions indicates that at that time—15.5 cal ka, climatic conditions were cold enough to support discontinuous permafrost.

Base-level Controls on Lake Hitchcock South of the Hartford Basin

Terminal Moraine, Block Channel spillway, and Block Delta

The story of glacial Lake Hitchcock actually begins with morainal deposition at the terminus of the late-Wisconsinan ice sheet between 24-25 cal ka. At that time, global sea level was 125 m below today's level and stood at the edge of the continental shelf, about 100 km south of the terminal position. The terminal moraine stands above today's sea level along much of its extent on Long Island, but lies below sea level between Montauk Point, L.I. and Block Island, R.I. and between Block Island and Martha's Vineyard (fig. 5). These submerged moraine sections and other glacial deposits and erosional features have been mapped using high-resolution seismic reflection profiles (Needel and others, 1983; Needel and Lewis, 1984). The submerged moraine crest lies generally at or above -20 m in altitude, and a deep notch is cut through the moraine 10 km west of Block Island. The bathymetric bottom of this notch is just below -30 m, but in the sub-bottom, the base of a buried channel lies at -70 m; this notch is called Block channel spillway (BCS) which served as the outlet for water spilling from glacial lakes in Block Island and Rhode Island Sounds. Another notch east of Block Island at the Mudhole (MH) was an earlier spillway for glacial Lake Rhode Island (Oakley and Boothroyd, 2013).

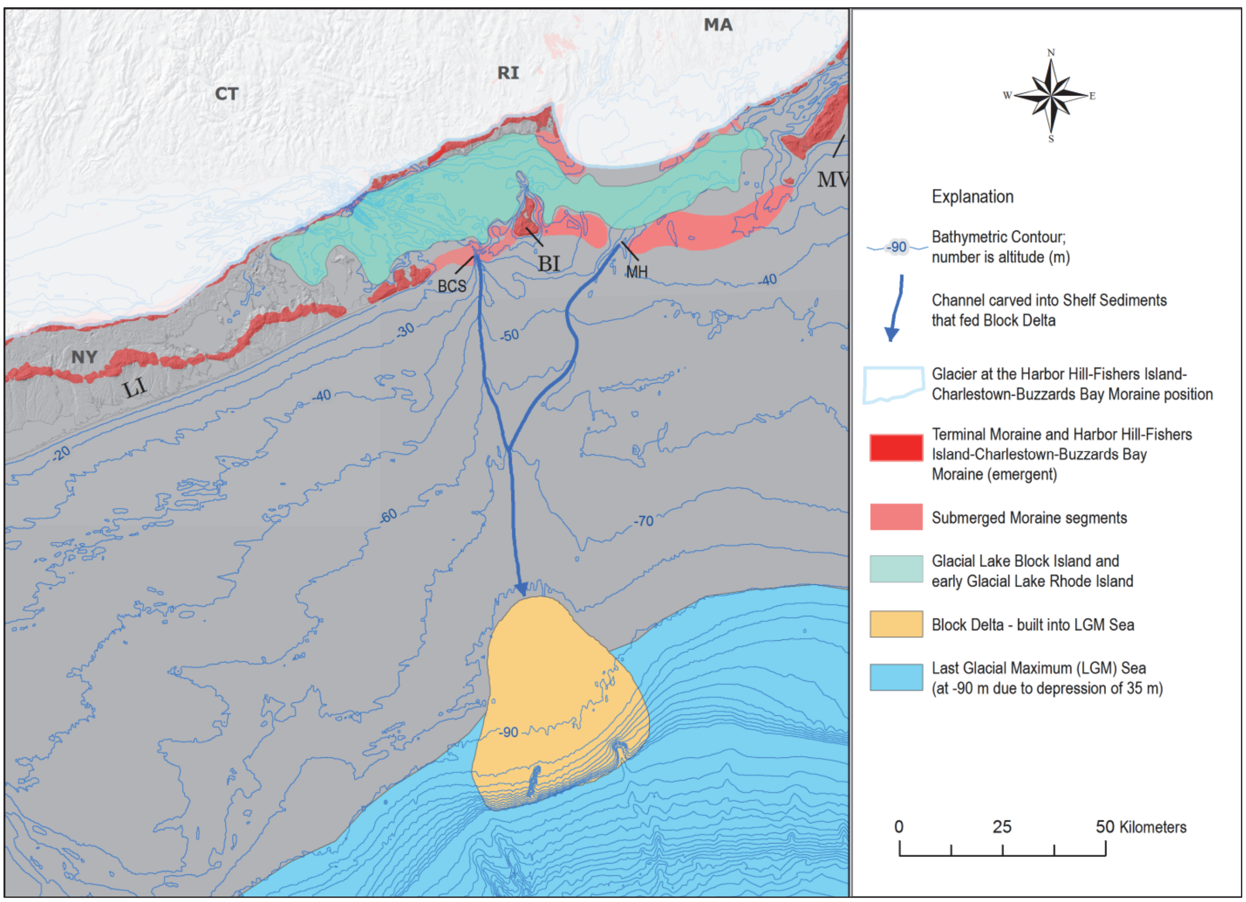


Figure 5. Map showing positions of Block delta, Terminal and Recessional moraines, glacial Lake Block Island and Rhode Island, Block channel spillway, and erosional channels cut into shelf sediments southward to Block delta. **LI**—Long Island, **BI**–Block Island, **MV**—Marthas Vineyard, **BCS**—Block Channel spillway, **MH**--Mudhole.

South of the spillway, a channel seen in the modern bathymetry leads to a broad surface at -80 to -90 m interpreted as a delta built into the LGM low-stand sea level by water draining through Block channel spillway notch (Garrison and McMaster, 1966). The break in slope at the front of the delta is at -90 m and records the water level into which the delta was built; it therefore represents a relative sea level at this location of -90 m. Since LGM eustatic sea level at that time was -125 m (Bard and others, 1990; Peltier, 2001), glacio-isostatic depression in the amount of 35 m is required to produce the -90 m relative level.

The terminal moraine sits atop Cretaceous coastal plain strata, and during its construction 24-25 cal ka (Rittenour and others, 2012, Balco and others, 2002; 2009; 2015), sea level stood at the edge of the Shelf about 115 km south of the ice position. Despite glacio-isosatic depression of about 50 m, the moraine ridge was constructed above sea level and the sea did not directly accompany the retreating ice northward.

Deposition of the terminal moraine began the process of glacial lake impoundment. The first lake behind the terminal moraine was glacial Lake Block Island (Oakley and Boothroyd, 2013). Successive ice-marginal lacustrine fans were deposited at grounding line positions on the lake bottom as the ice margin retreated northward through glacial Lake Block Island. Deltas built in front of the Harbor Hill-Fishers Island-Charlestown recessional moraine lie just off the Rhode Island coast at -10 to -20 m altitude (fig. 6). Delta altitudes were controlled by water spilling over the Block Channel spillway, initially at about -20 m. Extensive varved silt and clay drapes the ice-marginal fan deposits distal from the deltas in Block Island Sound. A deep channel (fig. 6) incised into the lake bed slopes southerly from about -50 m at the recessional moraine to about -70 m at BCS, indicating glacial lake drainage before transgression of the sea.

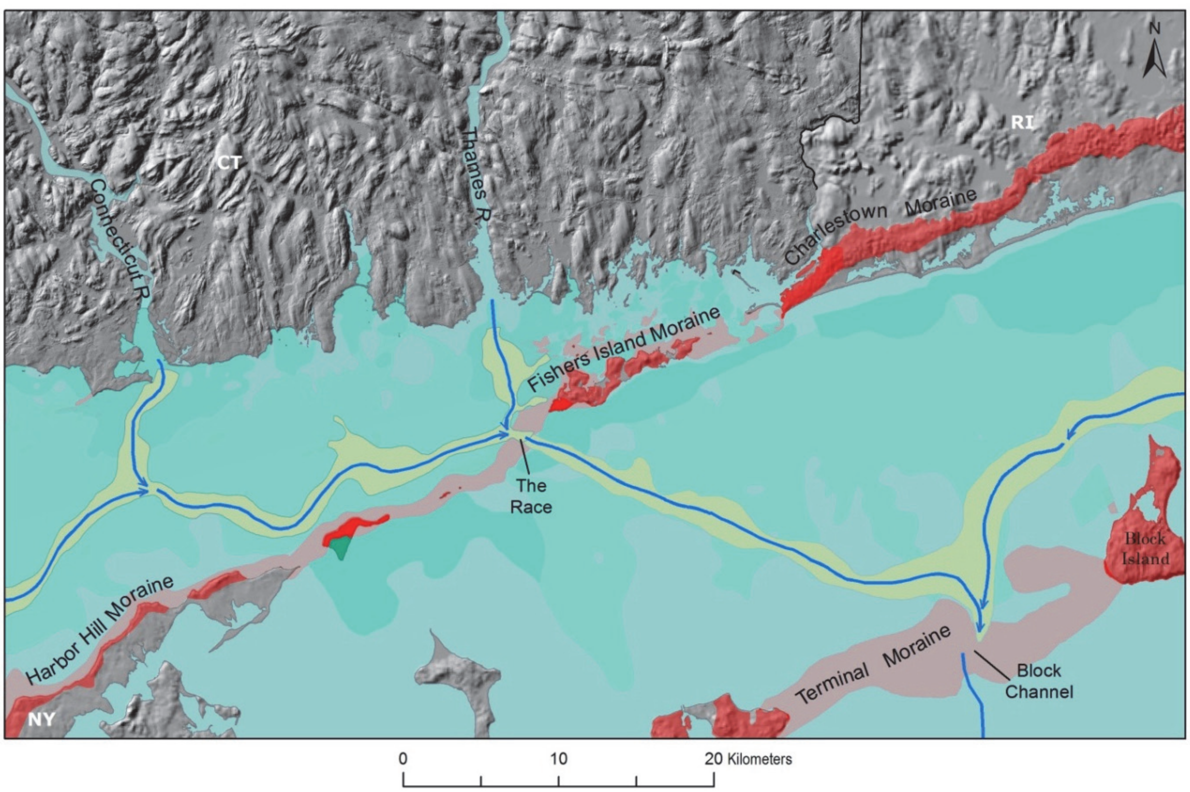


Figure 6. Map showing Harbor Hill - Fishers Island - Charlestown Moraine (red- emergent sections; light red-submerged sections) and position of The Race. Darker blue-green color indicates deltas of glacial Lake Block Island south of the moraine and glacial Lake Connecticut north of the moraine. Lighter blue-green color indicates lake-bottom sediment. Blue arrows on yellow polygon are channels cut into the lake beds.

The Race spillway at the Harbor Hill-Fishers Island-Charlestown Recessional Moraine

The recessional Harbor Hill – Fishers Island-Charlestown moraine was deposited at the ice margin about 21-22 cal ka. The crest of submerged sections of the recessional moraine lies at or above -10 m except at The Race 1.5 km west of Fishers Island (fig. 7), where a deep notch cuts the moraine down to about -60 m. Even deeper scour holes are present on both sides of the notch, exposing lake clay down to below -90 m. It is likely that some of the depth of the notch has been cut by modern scour because the Race is the location of daily tidal ebb and flow into Long Island Sound. During deglaciation, the notch was first the spillway for Glacial Lake Connecticut, and later, part of the pathway for meltwater escaping down the lower Connecticut River valley from glacial Lake Hitchcock. Initially, The Race spillway (TRS) was at about -10 m in altitude and controlled the level of the earliest deltas built into the lake in southeastern Connecticut. During the time of delta building, the spillway was eroded deeper into the moraine to at least -20 m as indicated by the level of the last delta built directly into the lake at New Haven. The spillway and lake level continued lowering as the ice retreated northward. A mapped sub-bottom channel system cutting the lakebed enters the spillway notch and emerges south of TRS at about -50 m heading to Block channel where it lies at -70 m. This channel was the pathway for water exiting the New Britain spillway from glacial Lake Hitchcock.

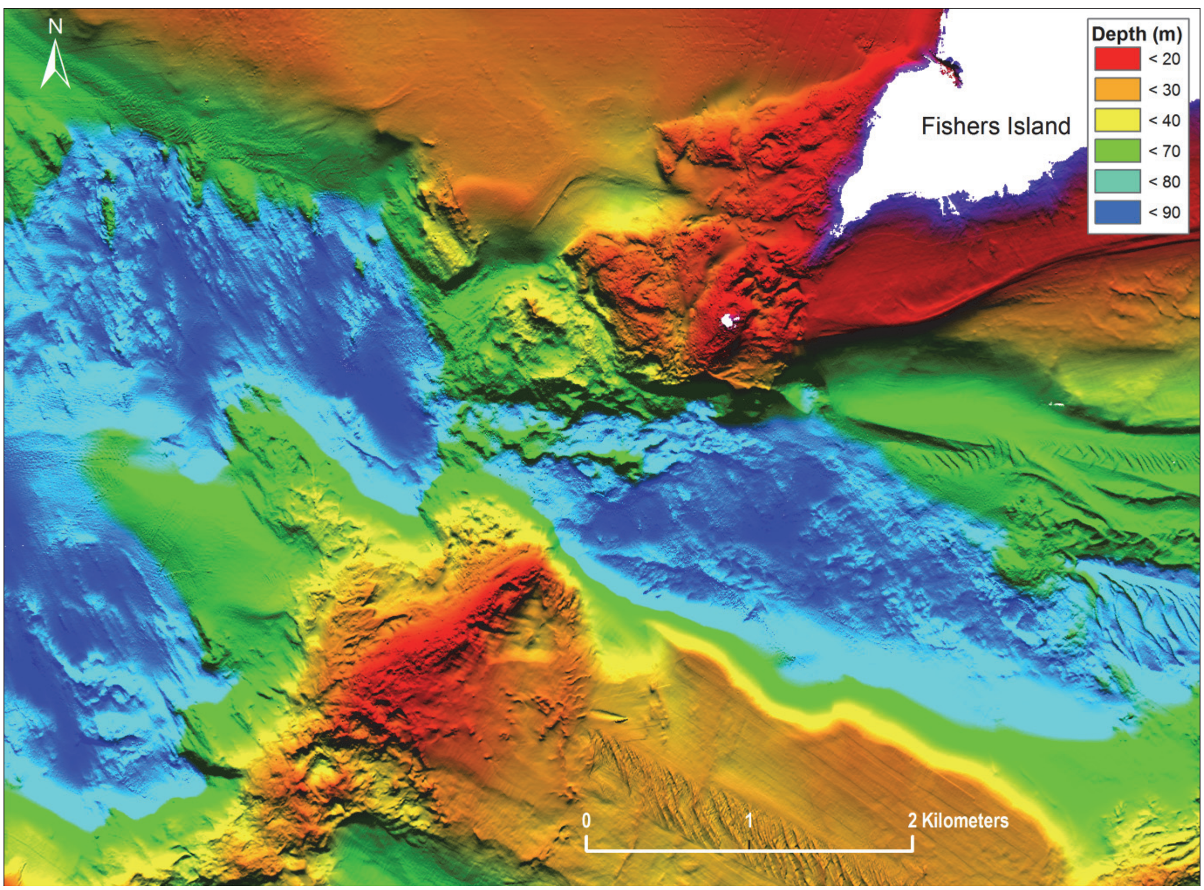


Figure 7. Multibeam sonar image of The Race spillway across the Harbor Hill - Fishers Island - Charlestown recessional moraine.

Long Island Sound Basin

Long Island Sound basin is an elongate E-W depression (fig. 8) formed by a south-sloping crystalline bedrock surface and a north sloping cuesta in Cretaceous Coastal Plain strata that overlie the bedrock surface from the middle of LIS southward. The basin is largely filled with deposits of glacial Lake Connecticut, including successive ice-marginal lacustrine fans, deltaic deposits, and thick lake-bottom varved silt and clay (fig. 9). The surface of these deposits became subaerially exposed as the lake lowered and drained, and a fluvial channel system cuts the drained lake bed (Lewis and Stone, 1991). A main channel extends from west to east across the Sound, and feeder channels that flowed north from Long Island and south from the Connecticut shoreline join the main channel (figure 8). Thalweg depths are shallower in the west and deepen to the east to about -50 m. In the easternmost part of LIS, much of the paleo-channel has been removed by modern tidal scour, but it must have exited via The Race spillway notch. The channel continues through Block Island Sound and through the notch at Block Channel spillway. The presence of this erosional channel system indicates that the glacial lakes in Block Island and Long Island Sounds had drained and the lake bottom was subaerially exposed before the sea was able to enter the basins. Only minor fluvial sediment is present in the lower part of the channel-fill deposits, and the remainder of the fill is estuarine mud. Very little sediment was being delivered to Long Island Sound by the Connecticut River during construction of the erosional channel, since most of it was being trapped in glacial Lake Hitchcock to the north.

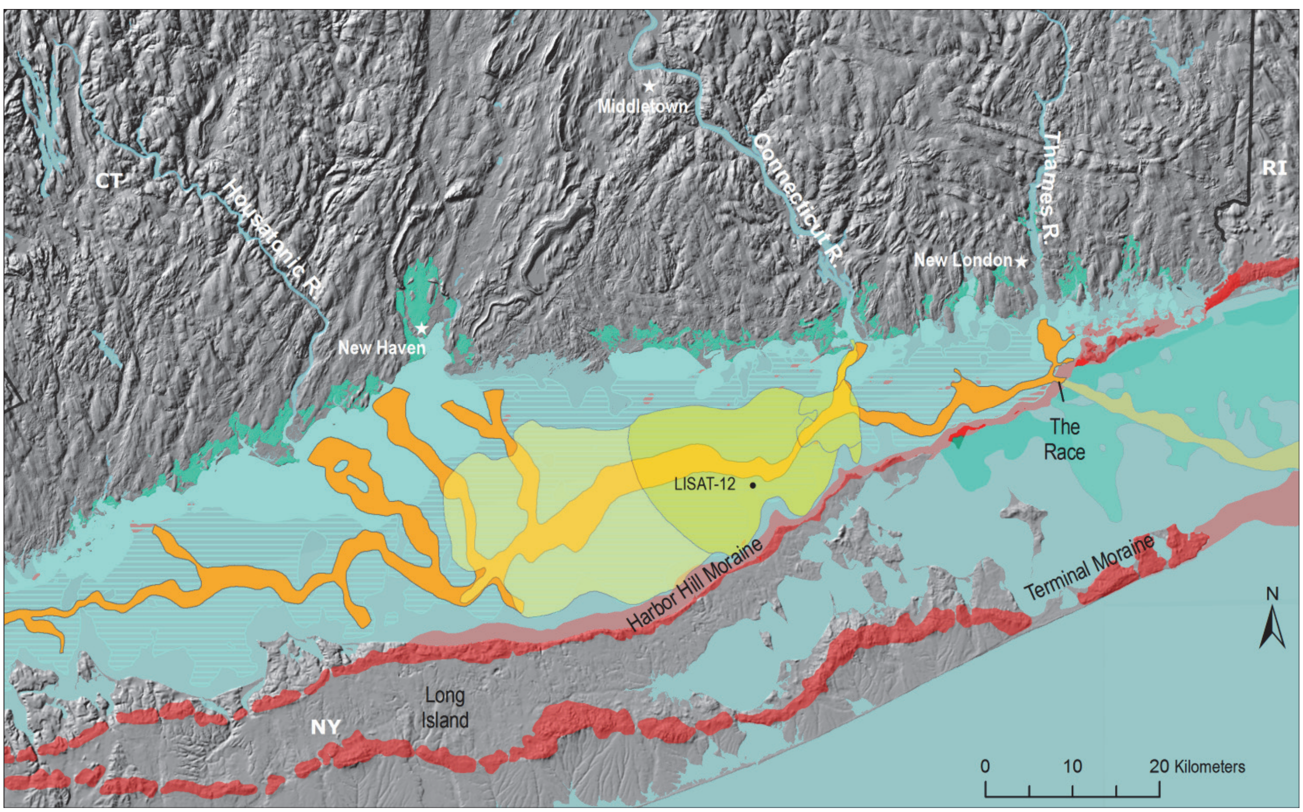


Figure 8. Map showing glacial and postglacial deposits in Long Island Sound. Sold blue-green—glacial Lake Connecticut deltas; line pattern blue-green—lake-bottom varved silt and clay. Red—emergent and submerged moraine segments; Orange—fluvial channel system. Yellow— -40-m marine delta-- darker shade is sandy foreset and topset beds, lighter shade is distal bottomset sediments. LISAT-12 is core location for radiocarbon dates.

A transgressive marine unconformity rising to as high as -20 m (Lewis and Stone, 1991) cuts the channel fill deposits and higher-lying lake beds (heavy solid black line on fig. 9). The sea was able to enter the Sounds via the erosional channel system as soon as down-cutting was completed and eustatic sea level began to rise above its LGM low-stand position (Lewis and DiGiacomo-Cohen, 2000). The Barbados sea level curve (Bard and others, 1990) indicates a slow eustatic rise of 0.44 m per century until 14.5 cal ka. By the time the stable level of glacial Lake Hitchcock was reached at around 17 cal ka, eustatic sea level had risen to -107 m, and due to glacio-isostatic depression, relative sea level In Long Island Sound stood at about -27 m.

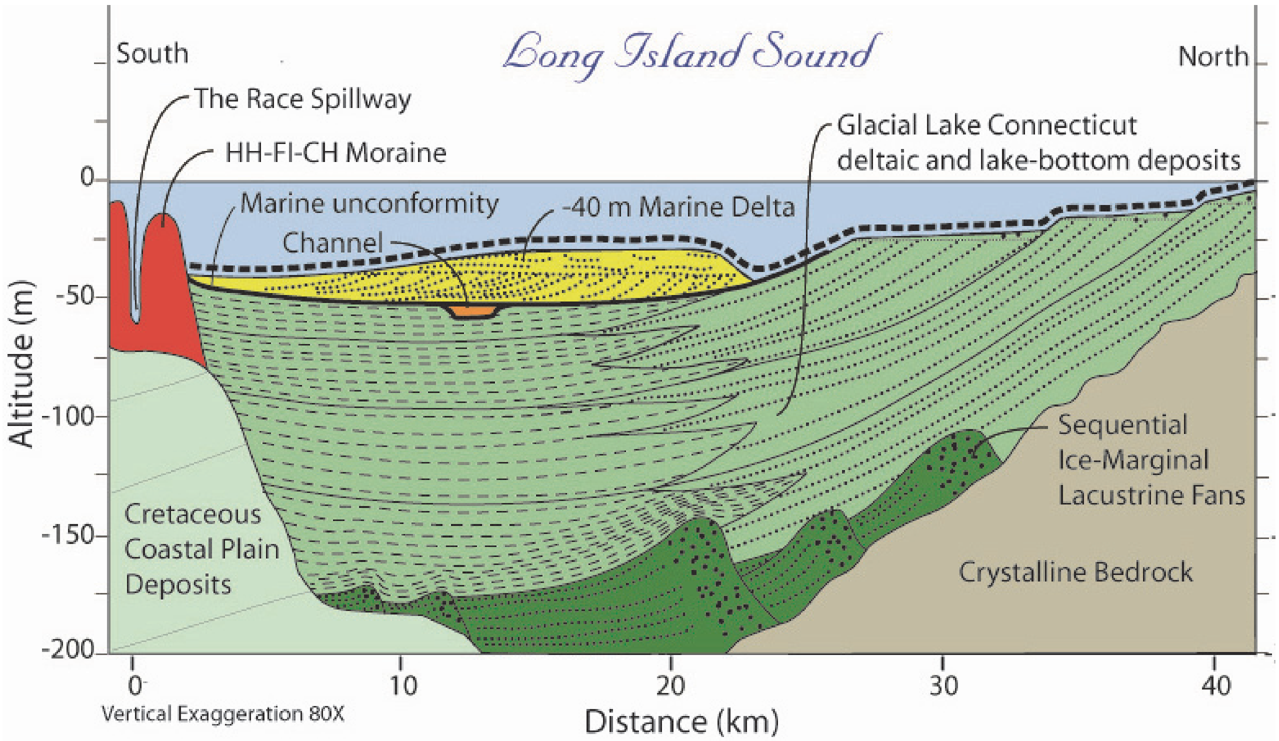


Figure 9. Schematic south to north geologic section across Long Island Sound. Upper heavy dashed black line represents Holocene marine mud of variable thickness that locally blankets older deposits. **HH**-Harbor Hill, **FI**-Fishers Island, **CH**-Charlestown.

An extensive delta sits atop the transgressive marine unconformity (figs. 8, 9), built westward from the mouth of the Connecticut River into the sea. The topset-foreset contact in the delta records a sea level stand of -40 m. The delta is approximately 750 km^2 in area, about 20 m thick in foreset sections, and contains more than 11 million m^3 of sediment. The source of this great volume of sediment was not available until the dam for glacial Lake Hitchcock was breeched and the Connecticut River began eroding the drained lakebed at about 15.6 cal ka, soon after the initiation of glacio-isostatic rebound. This extensive delta was constructed during the time of down-cutting of the drained bed of Lake Hitchcock along incised terraces (fig. 1) between 15.6 and about 9.0 cal ka. The volume of eroded lakebed sediment, as calculated from the area and depth of incised terraces, is 11-12 million m^3; this material now comprises the marine delta. The delta records a stable relative sea level of -40 m lasting for several thousand years. Figure 10 shows a conceptual relative sea level curve for Long Island Sound that is derived by combining the estimated amount of glacio-isostatic depression with the altitude of eustatic sea level from Bard and others (1990). The resulting curve shows that relative sea could have been at about -40 m between 15 and

11 cal ka because the rate of eustatic rise and the rate of glacio-isostatic uplift were nearly balanced. Several radiocarbon dates have been obtained on shells and other organic material in the marine delta. A ^{14}C date of 12,455$\pm$325(GX-18094) on detrital organic debris including shells reported by Lewis and others (1994) calibrates to 14.7 cal ka (Calib ver. 4.3) from the mid-delta section in core LISAT-12 provides maximum timeframe for delta construction in a -40-m sea. Several other ^{14}C dates from core LISAT-12 reported by Varekamp and others (2004) are on oyster shells (perhaps in bed-growth position) at -35 to -37 m in altitude; these have calibrated ages of 9.0 - 10.0 cal ka and provide a younger timeframe for a -40-m sea and delta construction.

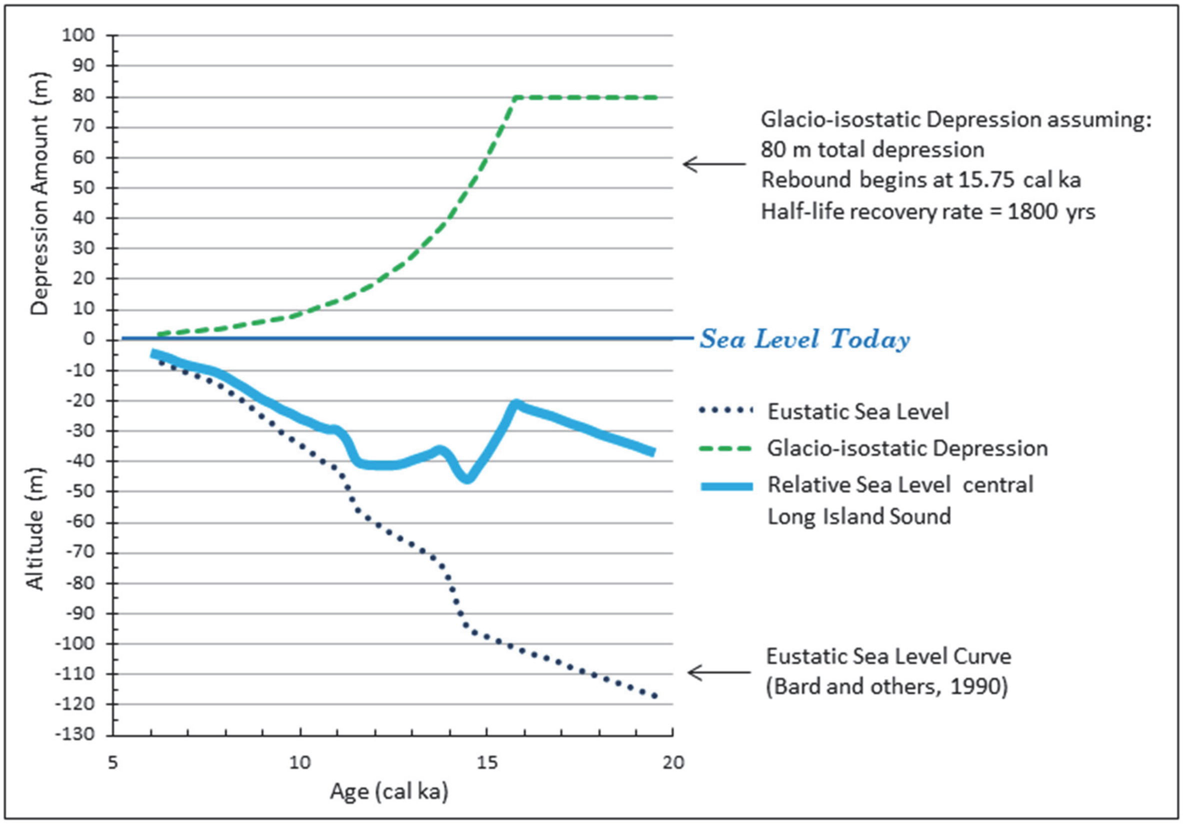


Figure 10. Conceptual Sea Level Curve for central Long Island Sound derived by combining estimated amounts of glacio-isostatic depression with altitudes of eustatic sea level from Bard and others (1990).

The sub-bottom features in Long Island Sound that are important to the Lake Hitchcock record are:

1) The part of the paleo-channel system that begins at the Connecticut River mouth and exits the Sound at the Race—this was the route for water spilling from glacial Lake Hitchcock.
2) The marine unconformity that cuts the channel-fill deposits and higher lying lake bed to as high as -20 m—providing evidence that the sea entered Long Island Sound and the lower Connecticut River estuary long before the dam for Lake Hitchcock was breeched.
3) The -40-m marine delta—the only source for the delta was the eroded bed of glacial Lake Hitchcock. The existence of this delta provides us with the means to estimate the amount of total depression in central LIS, and hence the rest of the western New England region; it also tells us the position of relative sea level for several thousand years after Lake Hitchcock drained.

Lower Connecticut River valley

The Connecticut River enters Long Island Sound between the towns of Old Saybrook to the west and Old Lyme to the east (fig. 11). The mouth of the River today is an estuary rimmed by salt marshes for about 12 km upriver to Essex; fresh-water tidal marshes flank the river northward for 40 km to Cromwell, and the River is tidal for another 35 km northward to Windsor Locks. At and north of Middletown, the river flows in the broad Hartford Basin that is underlain by Mesozoic rocks. From Middletown southward, the river leaves the Mesozoic Basin and is contained within a much narrower valley cut into Proterozoic and Paleozoic crystalline bedrock.

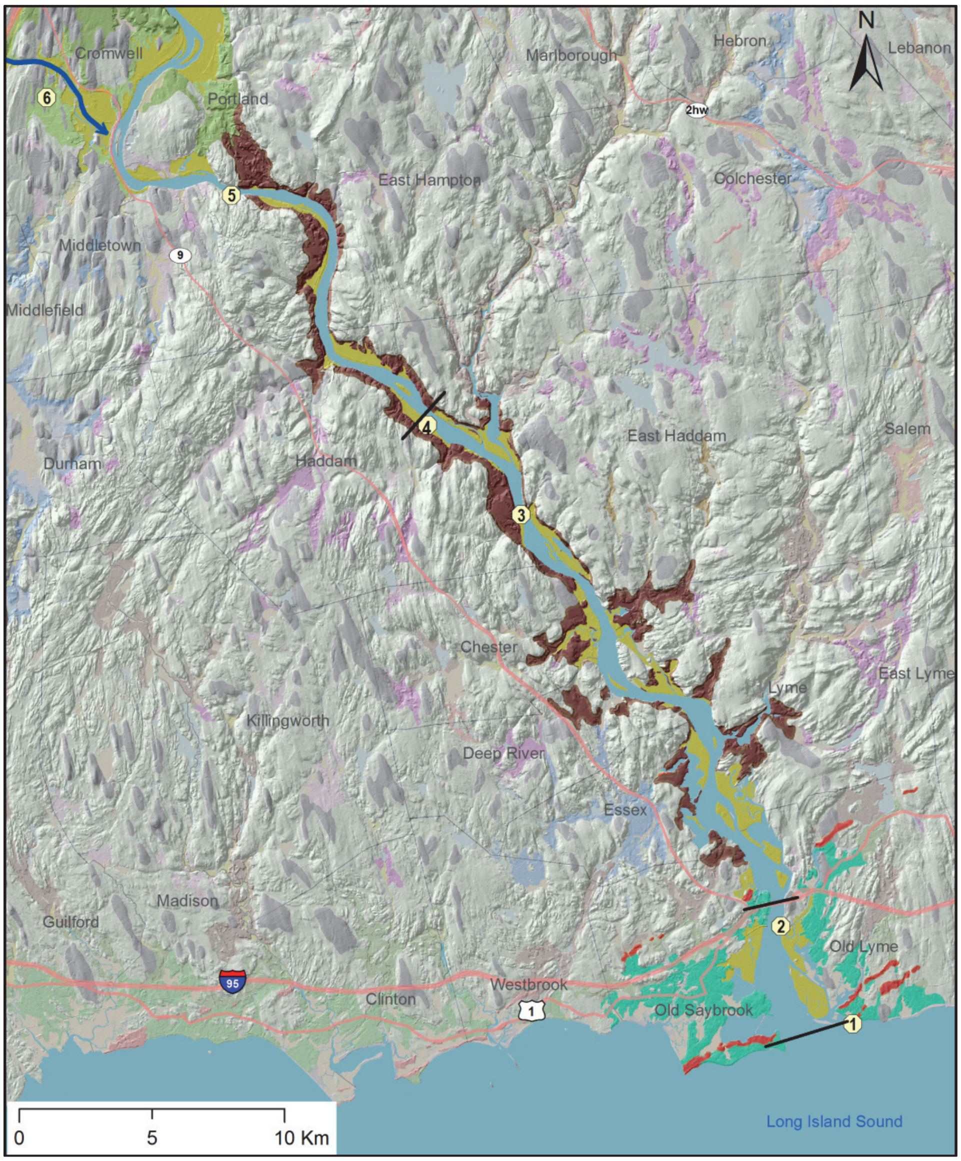


Figure 11. Glacial Lake deposits and moraines in the lower Connecticut River valley from Old Saybrook to Portland. Red indicates moraine deposits, bright green indicates deposits of glacial Lake Connecticut, brown is sediment-dammed pond deposits, dull green is glacial Lake Middletown deposits. Yellow is postglacial alluvium and salt-marsh deposits beneath which lies the channel pathway for water spilling from glacial Lake Hitchcock. Black lines are cross-section lines at Stops 1, 2, and 4. Numbered dots are Stop localities.

Moraines cross the lower Connecticut River at two places-- the Old Saybrook-Rocky Hollow-Wolf Rocks moraine at its mouth where it stands at the head of deltas built into glacial Lake Connecticut, and the Hammonasset-Ledyard moraine just north of the Baldwin Bridge (I-95). Meltwater deposits in the lower Connecticut River valley (fig. 11) are a 30-km long succession of ice-marginal deltas built into Glacial Lake Connecticut at the south end, into a series of sediment-dammed ponds throughout most of its length, and into Glacial Lake Middletown as the valley widens to the north. These deltaic deposits rise northward today from 3 m at Old Saybrook to about 60 m at Portland because they have been glacio-isostatically tilted. But when adjusted for that tilt, they reflect lowering water levels controlled by The Race and Block Channel spillways (figs. 6, 7).

Meltwater deposits in the lower Connecticut River valley served as a dam for glacial Lake Middletown and were incised by meltwater spilling from that lake. The rate of incision into these deposits was controlled by the rate of incision at TRS and BCS as well. By the time glacial Lake Hitchcock came into existence, the channel pathway through these deposits must have been incised down to at least 15 m altitude at The Straits (fig.13); otherwise water spilling from GLH would have been flowing uphill. Glacio-isostatically adjusted equivalent downstream altitudes on the incised channel pathway are -15 m at the mouth of the Connecticut River, -25 m at the Race, and -45 m at Block Channel; these downstream altitudes were likely deeper, since the channel pathway likely must have had a fluvial gradient of several m/km. By the time that the stable level of glacial Lake Hitchcock was reached at about 17.2 cal ka, altitudes of the pathway were at 15 m as it exited the NBS, 6 m at Middletown, -3 m at Haddam Meadows, -22 m at the Baldwin Bridge, -27 m at the mouth of the river, -50 m at the Race, and -70 m at Block Channel. When these altitudes are adjusted for differential glacio-isostatic depression, there is very little gradient on this pathway (fig. 12).

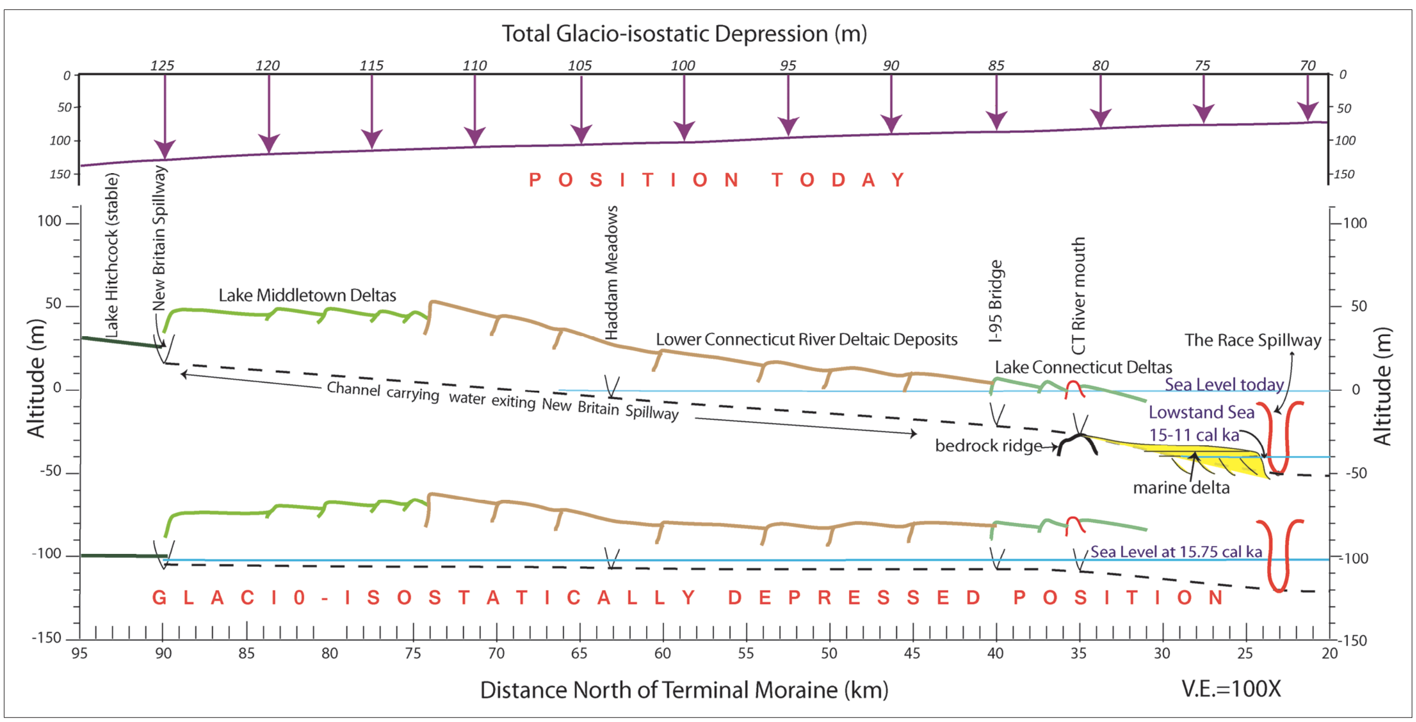


Figure 12. Profiles showing positions of delta deposits and other features in the lower Connecticut River valley and Long Island Sound. Dashed lines are positions of channel pathway for water spilling from glacial Lake Hitchcock at the stable level. Upper profile shows position of features today; lower profile shows glacio-isostatically depressed positions.

Glacial Lake Middletown

As the ice margin retreated northwestward from the Straits on the Connecticut River, the broader Middletown Basin was opened up and glacial Lake Middletown developed (fig. 13). The lower Connecticut River valley was filled with a long mass of successive ice marginal deltas built to altitudes above 45 m; this mass provided the dam for Lake Middletown, and water spilled over it, starting slightly below 45 m since the early Lake Middletown deltas at Portland (**Mp**) have topset-foreset contacts at about 45 m. Deltas in Lake Middletown are all at about the same level (45 m) despite the fact they lie successively farther north; when glacio-isostatic tilting is accounted for, we see that the lake level was lowering (fig. 12, lower profile) due to incision across the massive dam in the lower Connecticut River valley.

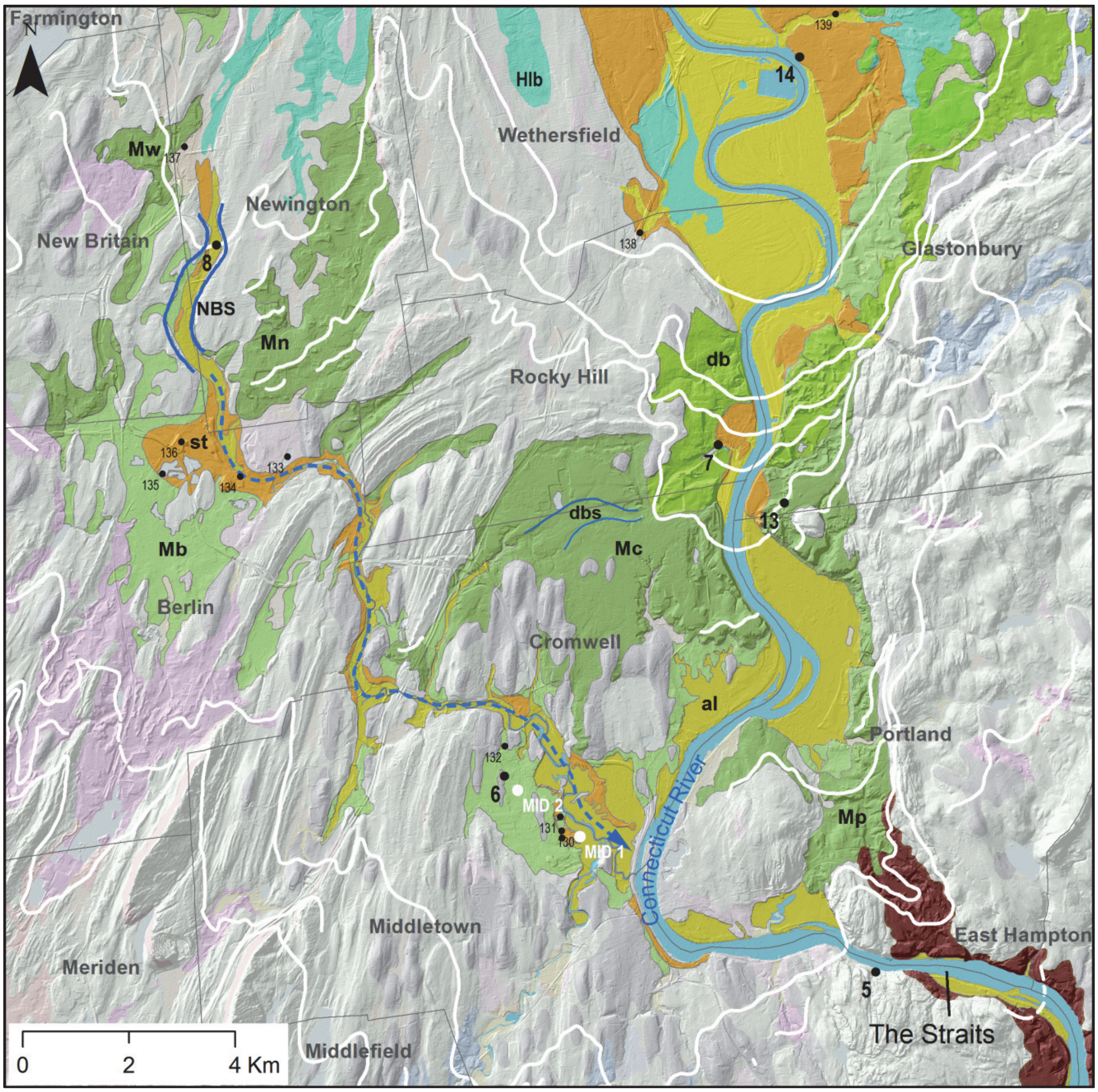


Figure 13. Deposits of glacial Lake Middletown modified from Stone and others, 2005**:** **Mp**—Portland deltas, **Mc**—Cromwell deltas, **Mn**—Newington deltas, **db**—Dividend Brook deltas, **Mb**—lake-bottom sediment, **Hlb**—Lake Hitchcock bottom deposits, **st**—stream terrace deposits, **al**—postglacial alluvium and swamp deposits, **dbs**—Dividend Brook spillway. White lines are retreatal ice-margin positions. White dots are varve core locations. Blue lines mark New Britain spillway channel for glacial Lake Hitchcock and dashed arrow marks pathway for water through the Lake Middletown basin along the present Mattabesett River to the Connecticut River.

Glacial Lake Middletown could alternatively be called "early glacial Lake Hitchcock." In fact, Lougee (1935; 1957) considered the Middletown basin to be part of Lake Hitchcock and its spillway to be over "compressed bouldery material caught in the West to East section of the Gorge (of the Connecticut River)" between Middletown and Haddam Neck (fig. 14).

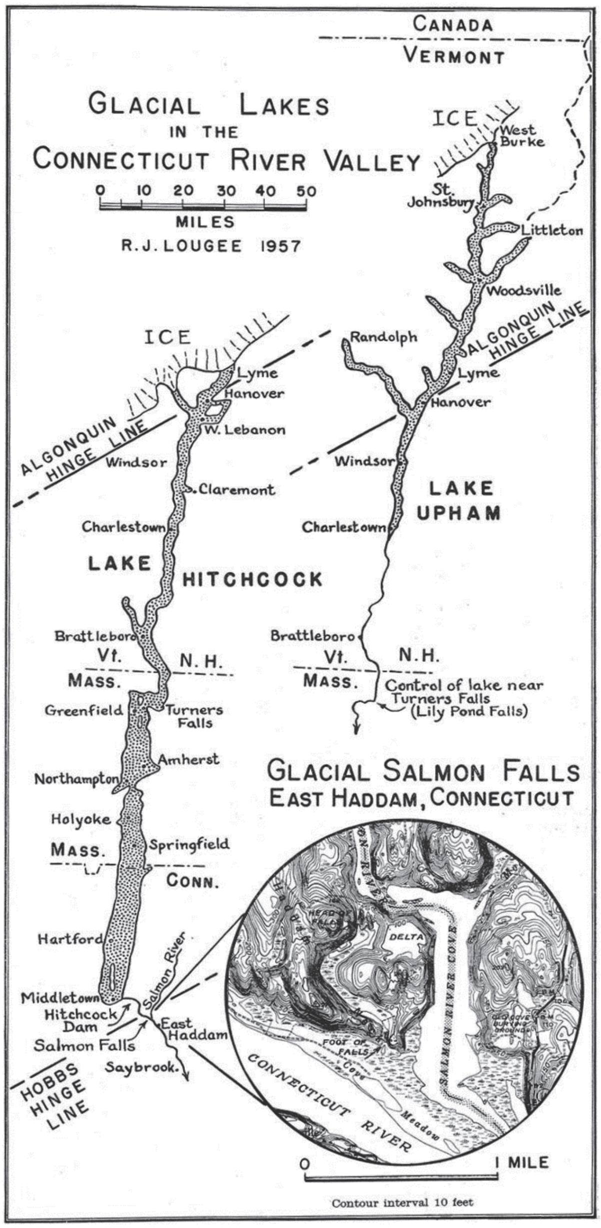


Figure 14. Diagram from Lougee, 1957, p. 27—shows Lake Hitchcock extending to Middletown and its dam occupying the Connecticut River gorge. Lougee believed that the presence of a spillway he termed glacial Salmon Falls (inset map) provided evidence for the location of the Lake Hitchcock dam.

Ice retreat from the narrower lower Connecticut River valley into the broad Mesozoic basin allowed lacustrine deposition to begin in a large open water body, where free-front deltas were built and lake-bottom plains formed. Deltas at Cromwell (**Mc**, fig. 13) were built freely into the lake; deltas to the north in Rocky Hill and Glastonbury (**db**) were ponded behind and water spilled across unit **Mc** at the Dividend Brook spillway (fig. 13). These deltas comprise the body of sediment that would later become the Rocky Hill dam for Lake Hitchcock. Deltas also formed in Newington (**Mn**) and New Britain (**Mw**). The water plane for Lake Middletown in the New Britain spillway area was about 48 m (fig. 3).

Lake-bottom plains (**Mb**) occupy low-lying areas through Berlin and Middletown. The entire lake-bottom section of Lake Middletown was sampled by side-by-side cores at 2 localities (MID1 & MID2, fig. 13); it contains a record of 171 years of varve deposition in the lake (see Stop 6 description). Varves likely accumulated in the Middletown basin only during the time that the ice margin stood in that part of the lake. So the ice-margin position at the head of unit **Mc** and extending westward covering the future location of the NBS (fig. 13) lasted for less than 200 yrs.

After further ice retreat and before Lake Hitchcock came into existence, Lake Middletown must have lowered by at least 10 m (to below 35 m). By the time Lake Hitchcock was at its stable level (25 m), Lake Middletown must have been lower than 25 m. The New Britain spillway partially cuts through deposits of Lake Middletown, and water exiting the spillway during the stable level carved into the lake-bottom sediments along the Mattabesset River (unit **st**, fig. 13). This terrace is preserved today at an altitude of about 15 m just south of the spillway and at about 6 m in Middletown.

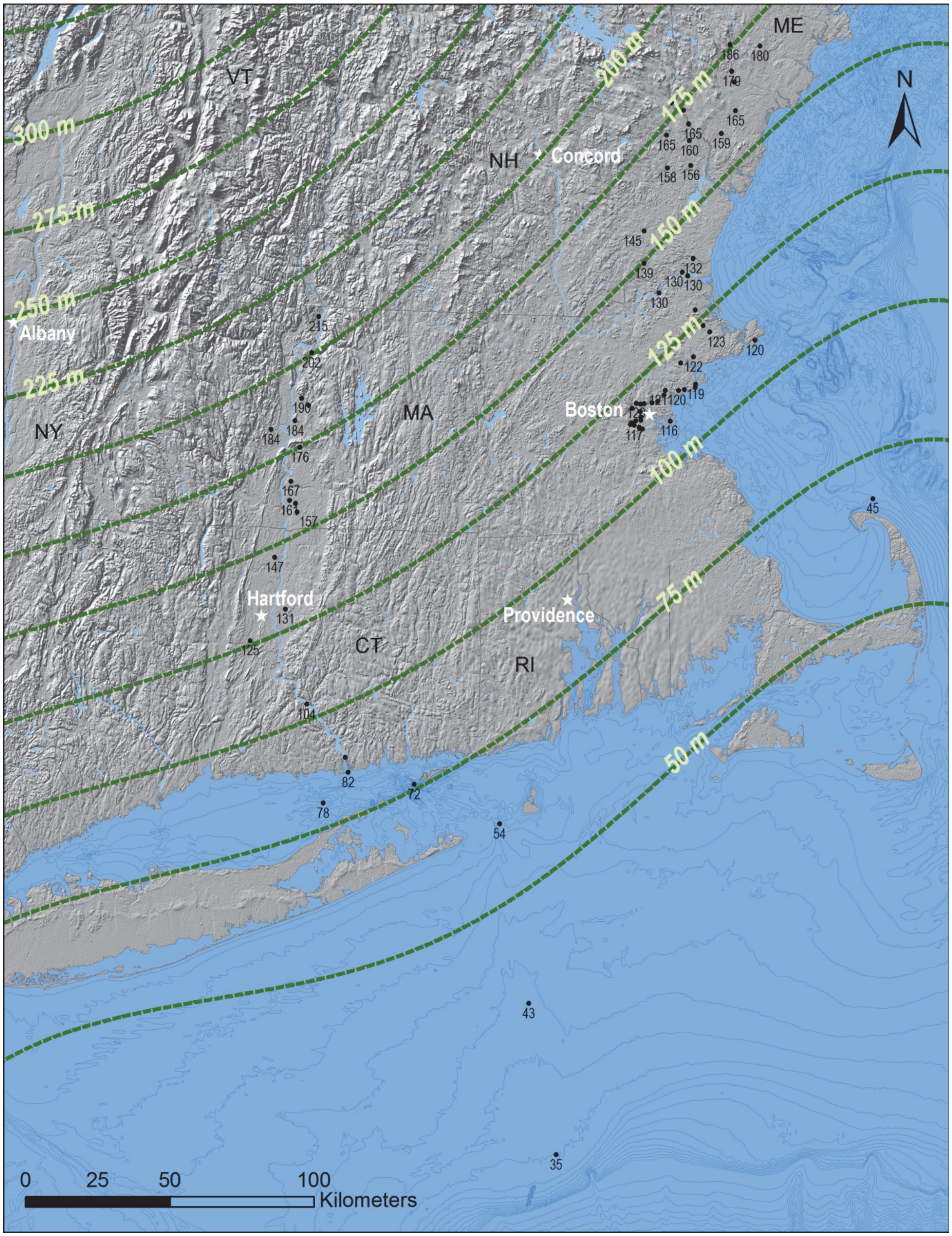


Figure 15. Total glacio-isostatic depression over the southern New England region. Data points indicate the amount of depression required to produce recorded synglacial sea levels in deltas along the coast from Boston to southern Maine and at Block delta, and total amount of depression at Lake Hitchcock spillway and delta sites, assuming 80 m in central Long Island Sound (requisite to produce a -40-m sea level as the lake drained). Surface generated using trend analysis in ARCmap.

Regional glacio-isostatic depression, lake levels, and relative sea level

The southern New England region was glacio-isostatically depressed during the late-Wisconsinan glaciation by increasing amounts to the north-northwest, the direction in which glacial ice was increasingly thicker. The amount of tilt and direction of tilting of this depression has been determined over the western part of the region by best-fit analysis on glacial Lake Hitchcock delta altitudes that rise 0.9 m/km in a N21°W direction (Koteff and Larsen, 1989). Similar tilt directions and slopes have been determined from deltas of other smaller glacial lakes and glaciomarine deltas across the region (Koteff and Larsen, 1989; Koteff and others, 1993; Stone and others, 2005; Oakley and Boothroyd, 2012).

Since detailed eustatic sea level data became available in the late 1980s and early 1990s (Fairbanks, 1989; Bard and others, 1990; 1993; Peltier, 2001; Peltier and Fairbanks, 2006), it has become possible to determine the total amount of depression needed to produce relative synglacial sea levels during deglaciation; these relative levels were above today's sea level from Boston northward in Massachusetts, coastal New Hampshire, and southern Maine. Total depression amounts for the non-marine features to the west is calculated using the 80 m of total depression in central Long Island Sound necessary to produce the -40 m sea level recorded by the marine delta (see figs. 7, 8, and 9). Figure 15 shows depression isobases for the southern New England region based on these requisite depression amounts.

The New Britain spillway was not an independent control for GLH levels; deepening of the spillway channel from higher levels to the stable level was controlled by conditions to the south. The upper profile in figure 16 shows the positions of Lake Hitchcock deltas and other important features to the south as they exist today; the lower profile shows their positions when glacio-isostatic depression is restored. As can be seen on the lower profile, stable-level deltas of glacial Lake Hitchcock constructed between 17.3 and 15.6 cal ka were at -100 m altitude in their depressed state. In order for the lake to be at that level, the notch spillways through the terminal and recessional moraines and deltaic deposits in the lower Connecticut River valley must have been entrenched to at least that level-- a depth of 30-50 m below their original surface altitudes. The dashed black line on the profiles represents the base of the channel pathway for water exiting the New Britain spillway at the stable-lake-level of 25 m (-100 m in depressed position). Altitudes of this channel today are 6 m at Middletown, -3 m at Haddam Meadows, -22 m at the Baldwin Bridge, -27 m at the mouth of the Connecticut River, -50 m at the Race, and -70 m at Block Channel. In their depressed position, these altitudes were -110 m at Middletown, -107 m at Haddam Meadows, -108 m at the Baldwin Bridge, -109 m at the mouth of the River, -122 m at the Race, and -124 m at Block Channel—just above LGM eustatic sea level at -125 m. This was as deep as entrenchment could go because of base sea level at the time. The New Britain spillway could not have lowered further than the 25-m level for at least 2 reasons:

1. The base of the channel through which the paleo-Connecticut River carried water spilling from Lake Hitchcock was imposed on bedrock at -27 m altitude at the mouth of the Connecticut River (-109 m in depressed position); therefore no more down-cutting could take place.
2. LGM low-stand eustatic sea level had risen to -109 m by as early as 17.5 cal ka (Bard and others, 1990; Peltier and Fairbanks, 2006), so the sea was already making its way up the Connecticut River estuary. By 16.0 cal ka, sea level stood at -102 m, and with a couple more meters of rise would have flooded the New Britain spillway (fig. 16).

Flooding of Lake Hitchcock by the Sea did not take place because glacio-isostatic rebound began at about 15.7 cal ka. The lake was able to lower below the stable stage only after the Rocky Hill dam was breeched, the lower Connecticut valley was being uplifted, and the sea was receding to lower levels.

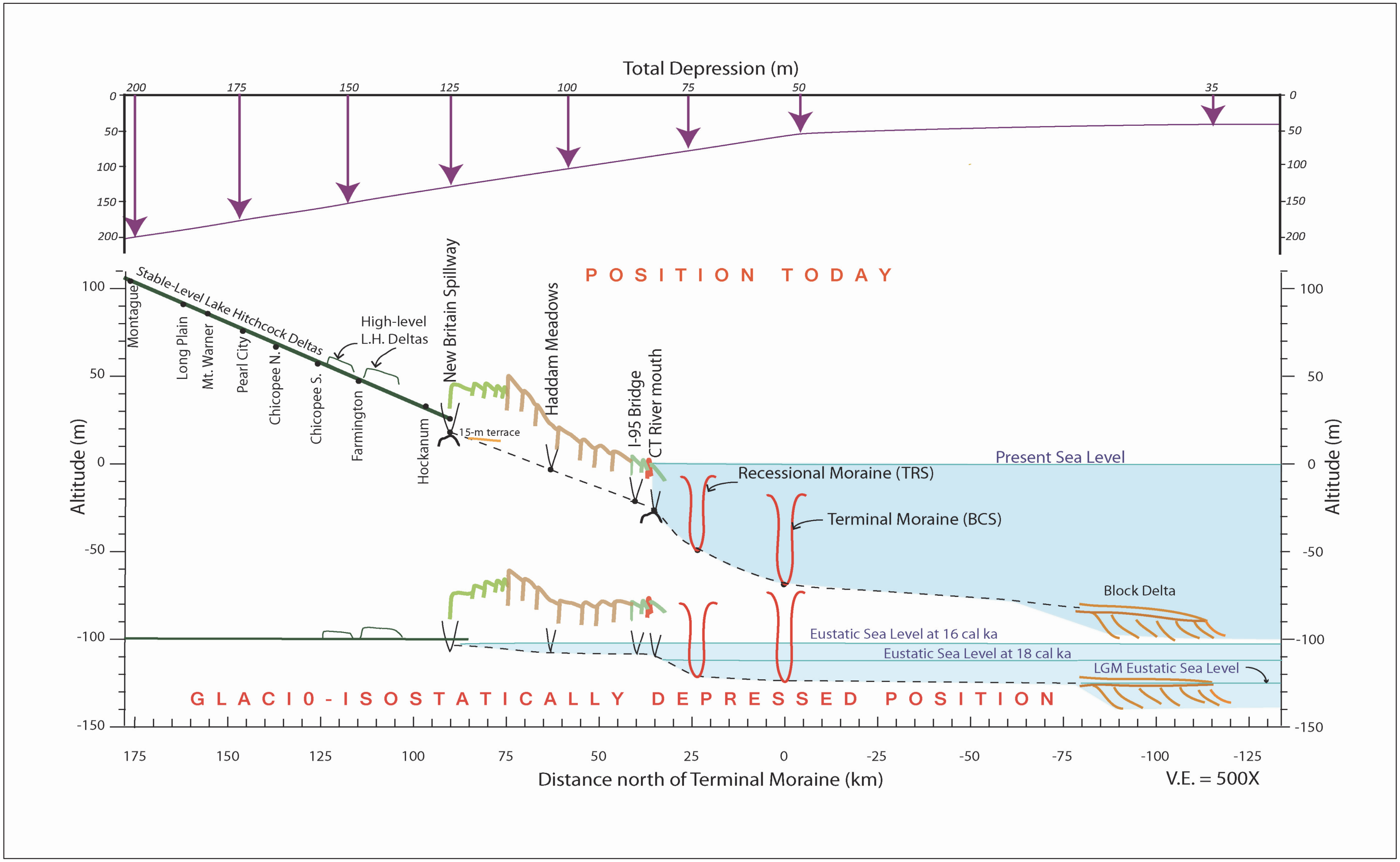


Figure 16. Amount of glacio-isostatic depression shown by purple line. Upper profile is the present position of Lake Hitchcock deltas, Lake Middletown deltas (light green), lower Connecticut River deltaic deposits (light brown) and the channel that cuts them, The Race spillway (TRS), Block Channel spillway (BCS), and Block delta. Lower profile is the glacio-isostatically depressed positions before rebound began.

The dam was most likely breached 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, or 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 (**Hpsf**). Once this process began, it proceeded rapidly as the dam was incised from just above 18 m in altitude (the stable level at the dam) to just above 12 m; once this 6 m of lowering was accomplished, glacial Lake Hitchcock, south of the Holyoke Range, was entirely drained and the newly formed Connecticut River began to incise the lake floor over the 80-km stretch between the Holyoke Range and the breached dam. Glacial Lake Hitchcock continued to exist north of the Holyoke Range with initial water depths of about 40 m (lowered from stable level by only 6 m). Continued lowering of the lake was controlled by the rate of rebound, which made it possible for the lakebed south of the Holyoke Range to be incised. The drained lakebed provided a major new sediment source for the Connecticut River that continued to flow southward to Long Island Sound where sea level then stood at about -40 m for several thousand years while the rate of glacio-isostatic uplift nearly matched the rate of sea level rise during meltwater pulse 1A and the Younger Dryas cold-climate event (fig. 10). The extensive marine delta in Long Island Sound (fig. 8) was constructed during this time of relatively stable sea level.

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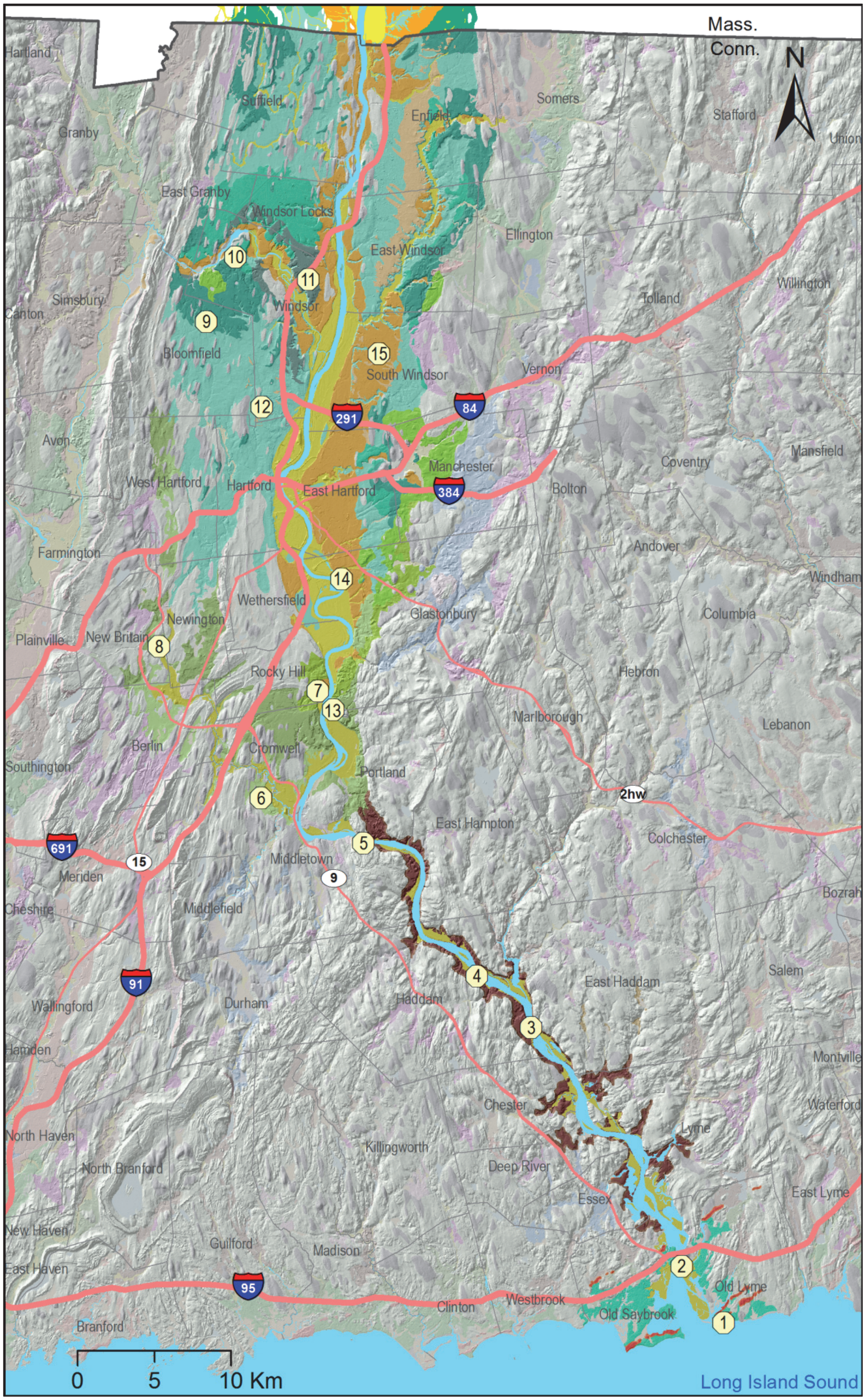
Fieldtrip Stops



Stop 1. Griswold Point, Old Lyme (-72.313 41.278). Griswold Point is on land owned by the Nature Conservancy; however, the overland access road crosses private property. Access is only possible through permission of the landowner.

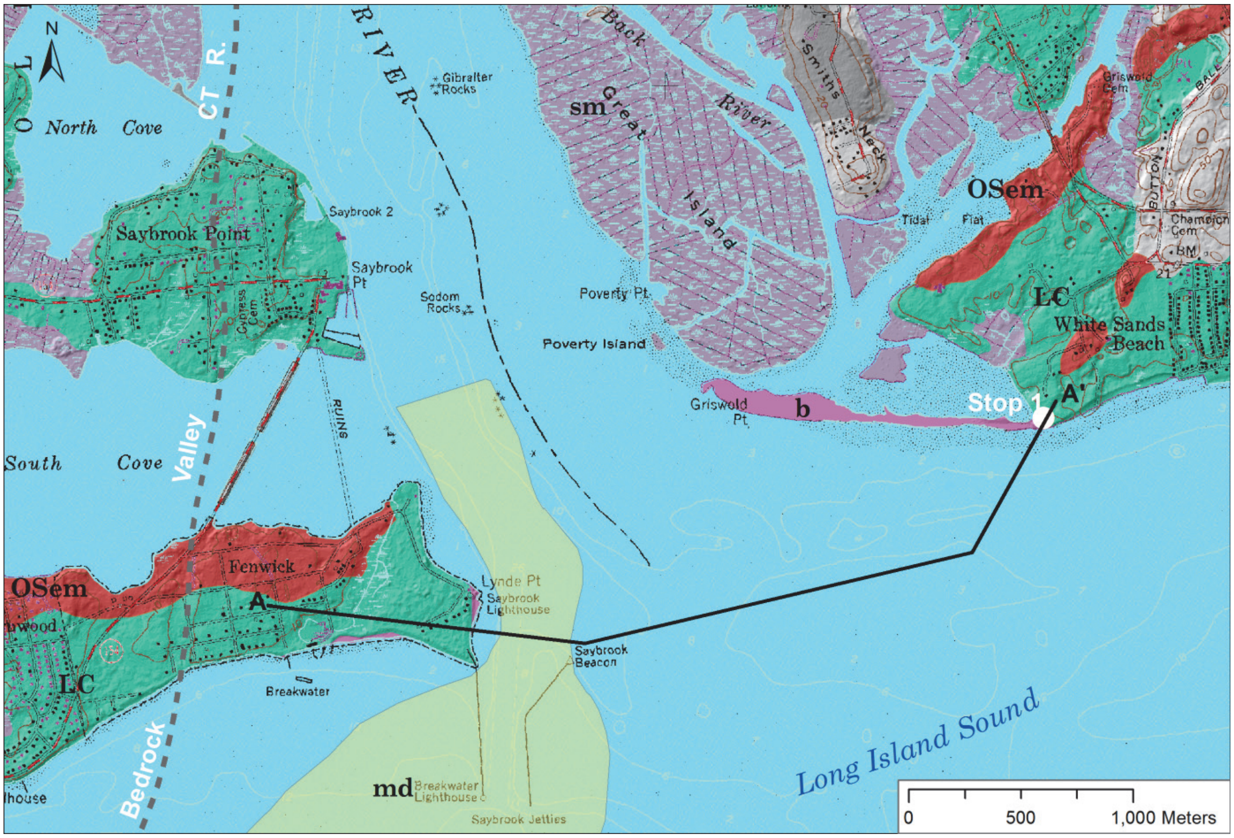


Figure 17. Quaternary geologic units in the vicinity of Stop 1 modified from Stone and others, 2005. **LC**—ice-marginal deltas of glacial Lake Connecticut; **OSem**—Old Saybrook moraine; **md**—marine delta; **sm**—salt-marsh deposits; **b**—beach and spit deposits. A-A' is section line (fig. 19).

Griswold Point, as labeled on the topographic map (fig. 17) is at the western end of a spit that is today no longer connected to the mainland and can only be reached by water. We will stop at the wave-cut front of an ice-marginal delta built into glacial Lake Connecticut in Long Island Sound. A segment of the Old Saybrook moraine lies at the northern edge of the delta. The sea cliff at times exposes topset and foreset beds of the delta, which record an approximate 1-m-altitude lake level. Our main purpose at this stop is to discuss the geology here beneath the River and Long Island Sound that is important to understanding the levels and drainage of glacial Lake Hitchcock to the north. Looking west, we can see the Saybrook lighthouse at Lynde Point on the other side of the River that also occupies the surface of a delta in front of the Old Saybrook moraine. Delta surfaces here at the mouth of the River are 3-4 m in altitude. The deep bedrock valley of the Connecticut River lies beneath the Saybrook Point and Fenwick peninsulas, and can be seen just offshore on seismic-reflection profiles to be deeper than -125 m in altitude. The bedrock valley is filled with deposits of glacial Lake Connecticut.

Figure 18. Sea cliff at Stop 1.

At Stop 1 we are at the mouth of the Connecticut River where it meets Long Island Sound. Cross-section A-A' (fig. 19) constructed using high-resolution seismic reflection profiles offshore shows a channel incised into glacial Lake Connecticut deposits with its base superimposed on bedrock at -27 m (-109 when glacio-isostatically depressed). The base of this channel is the lowest entrenched position of water flowing from the New Britain spillway down the Connecticut River valley. Fluvial sediment that partially fills the channel is material eroded from the drained bed of Lake Hitchcock and is on grade to the -40-m marine delta in Long Island Sound (fig. 17, fig. 8).

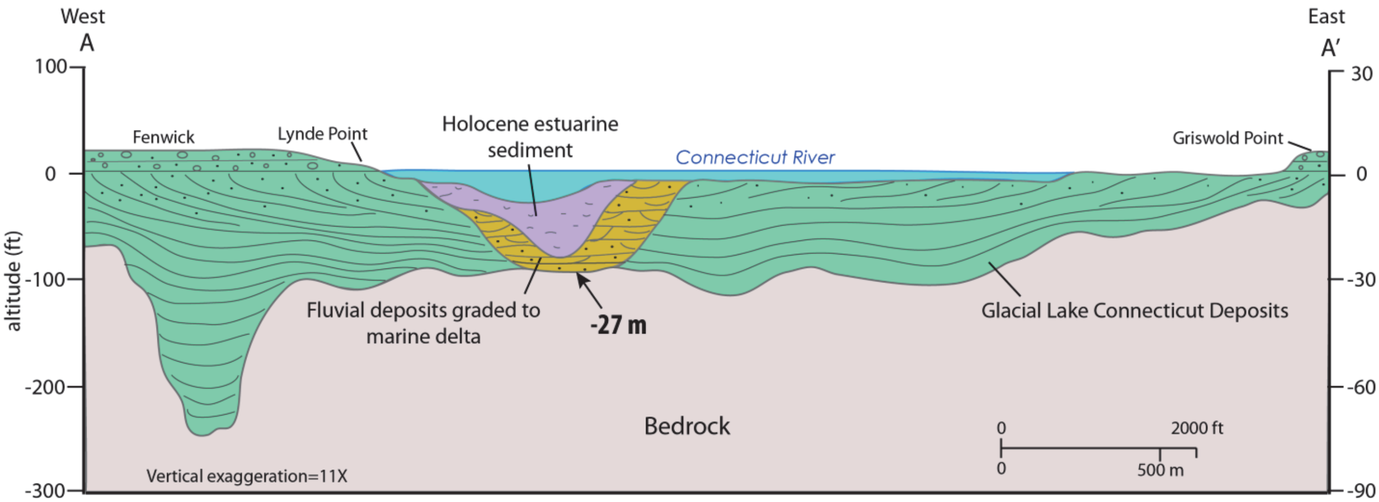


Figure 19. Cross-section A-A' at the Connecticut River mouth where it meets Long Island Sound. See line of section--figure 17.

Enroute to Stop 2. Along Rt. 156 (Shore Rd.) we cross the kettled surface of deltaic deposits, then rise slightly onto a boulder ridge that is the Old Saybrook moraine. Overall this moraine trends ENE-WSW and continues eastward in segments to the Wolf Rocks moraine in North Kington, RI. It goes offshore west of the Connecticut River at Cornfield Pt. in Old Saybrook. After crossing the Black Hall River on Rt. 156, we continue traveling over deltaic surfaces of glacial Lake Connecticut until the left turn for Stop 2 after crossing the Lieutenant River.

Stop 2. DEEP marine headquarters, 333 Ferry Road, Old Lyme. (-72.346, 41.311)

Looking north from this vantage point we have an excellent view of the Baldwin Bridge (I-95) (fig. 20), which crosses the Connecticut River at its narrowest point in the lower estuary region. We are on the East side of the River underlain here by till and bedrock (fig. 21). On the West side there are ice-marginal deltaic deposits of Lake Connecticut. Delta plain surfaces here are at 7-10 m in altitude with a topset-foreset contact at about 6 m. Cross-section B-B' (fig. 22) was constructed using bridge-boring data for the new span that was built in 1989-90, excavations on both sides of the river during construction, other well data, and seismic-reflection profiles.

Figure 20. North-northwesterly oblique aerial view of the Baldwin Bridge (Rt. I-95) spanning the Connecticut River.

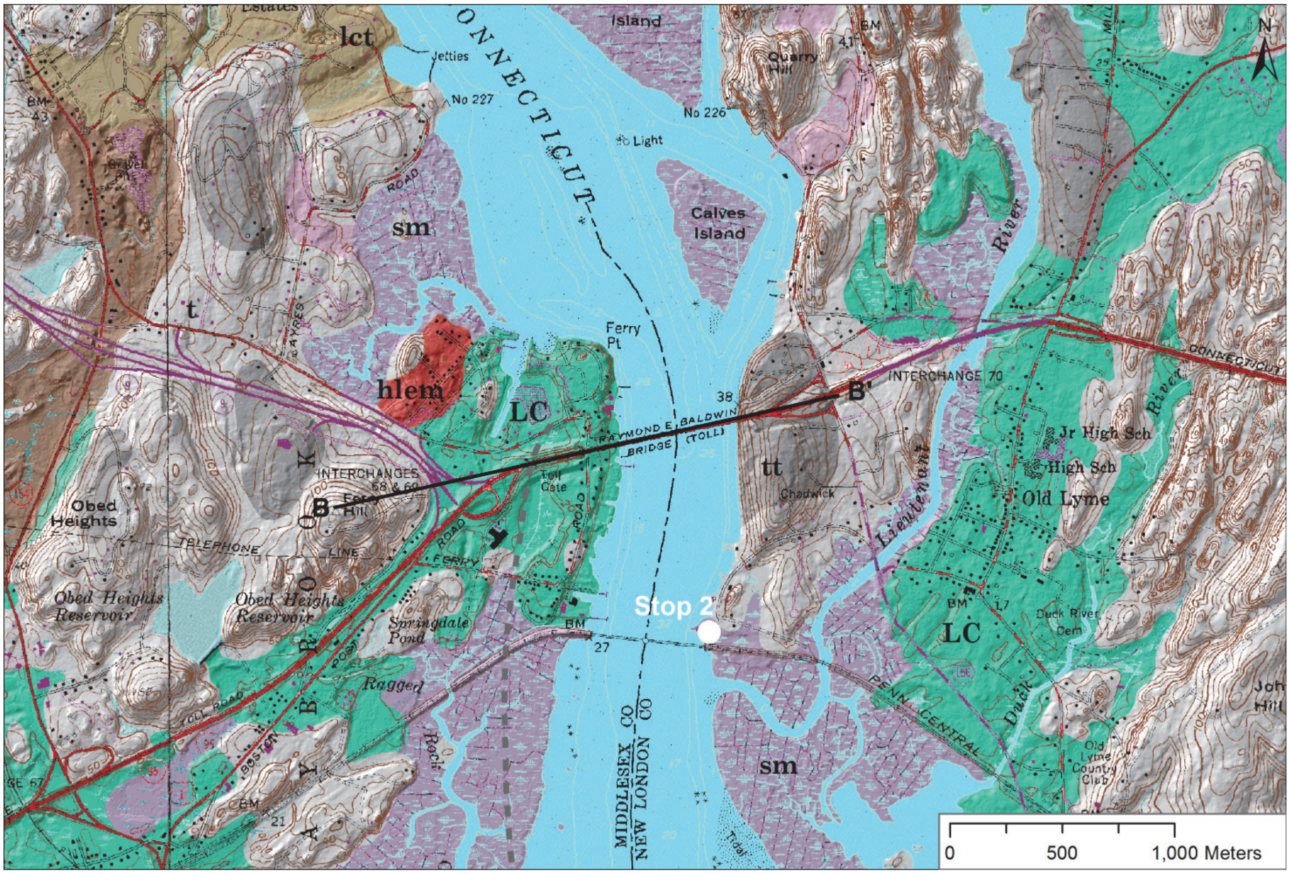


Figure 21. Quaternary geologic units in the vicinity of Stop 2 modified from Stone and others, 2005. **LC**—deposits of glacial Lake Connecticut, **lct**—lower Connecticut River deltaic deposits, **hlem**—Hammonasset-Ledyard moraine deposits, **t**— till deposits, **tt**—thick till deposits, **sm**—salt-marsh and estuarine deposits. B-B'—line of section (fig. 21).

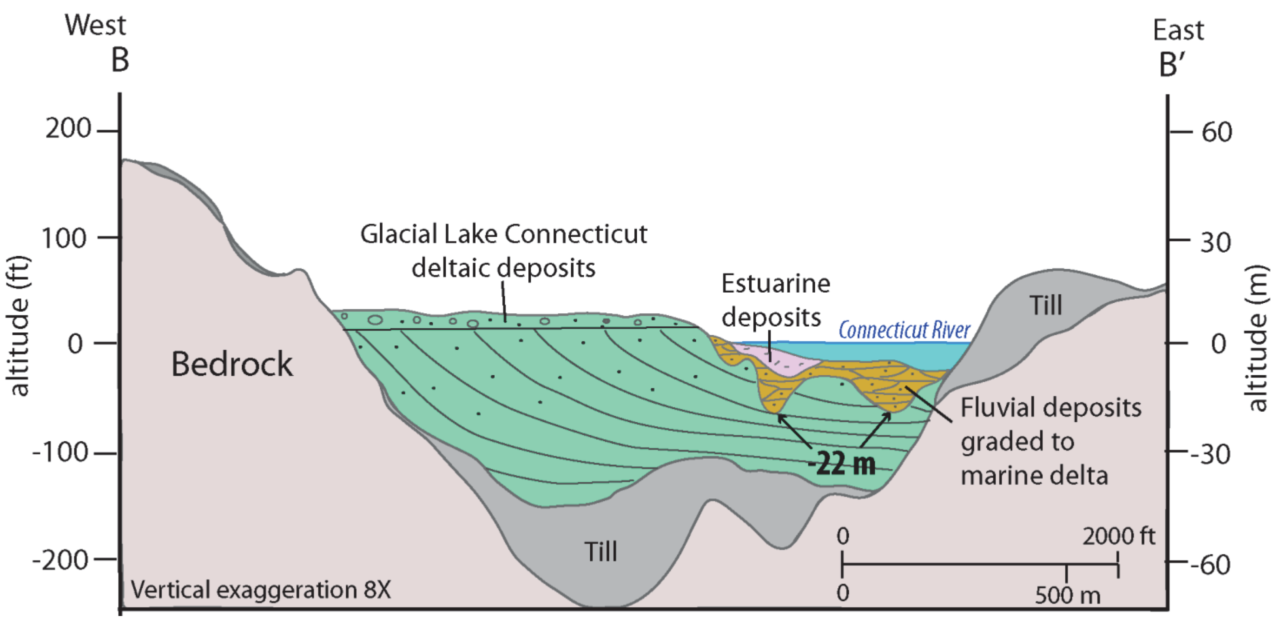


Figure 22. Cross section B-B' at the Baldwin Bridge, Old Saybrook to Old Lyme. Constructed using detailed bridge-boring data for the new bridge and water-well data.

Bridge-boring logs indicate that Lake Connecticut deltaic deposits consist of gravelly, coarse to fine sand topset beds that overlie fine to medium sand foreset beds and fine sand and silt bottomset beds; fluvial deposits graded to the marine delta are medium sand and organic silt; estuarine deposits consist of soft organic silty clay. The most important feature shown on the cross section to the Lake Hitchcock story is the channel incised into glacial Lake Connecticut deposits. This channel is the upstream equivalent of the one shown on the river mouth section (A-A' fig. 19) The base of the channel lies at -22 m here, 5 m higher than at the mouth of the River; but here we are 5.5 km farther north where there was more depression—the depressed altitude here is -108 m, only 1 m higher than at the mouth. This channel first carried water spilling from glacial Lake Middletown, then water spilling from the New Britain spillway during the high-level and stable-level stages of glacial Lake Hitchcock. By the time that Lake Hitchcock reached the stable-level stage, the channel could not be further incised because it was superimposed on bedrock at the mouth of the River. Water continued to follow this channel after the Rocky Hill dam was breeched and Lake Hitchcock lowered exposing its bed. The lakebed was incised as rebound progressed, and the channel became the pathway for sediment being carried to the marine delta in Long Island Sound.

After discussion of bridge cross section here, we will board the River Quest expedition vessel (fig. 23) in order to view the geology from the vantage point of the River for about 20 km upstream to the next bridge.

Figure 23. River Quest vessel departing home base. Historic Haddam-East Haddam Swing Bridge and Goodspeed Opera House in background.

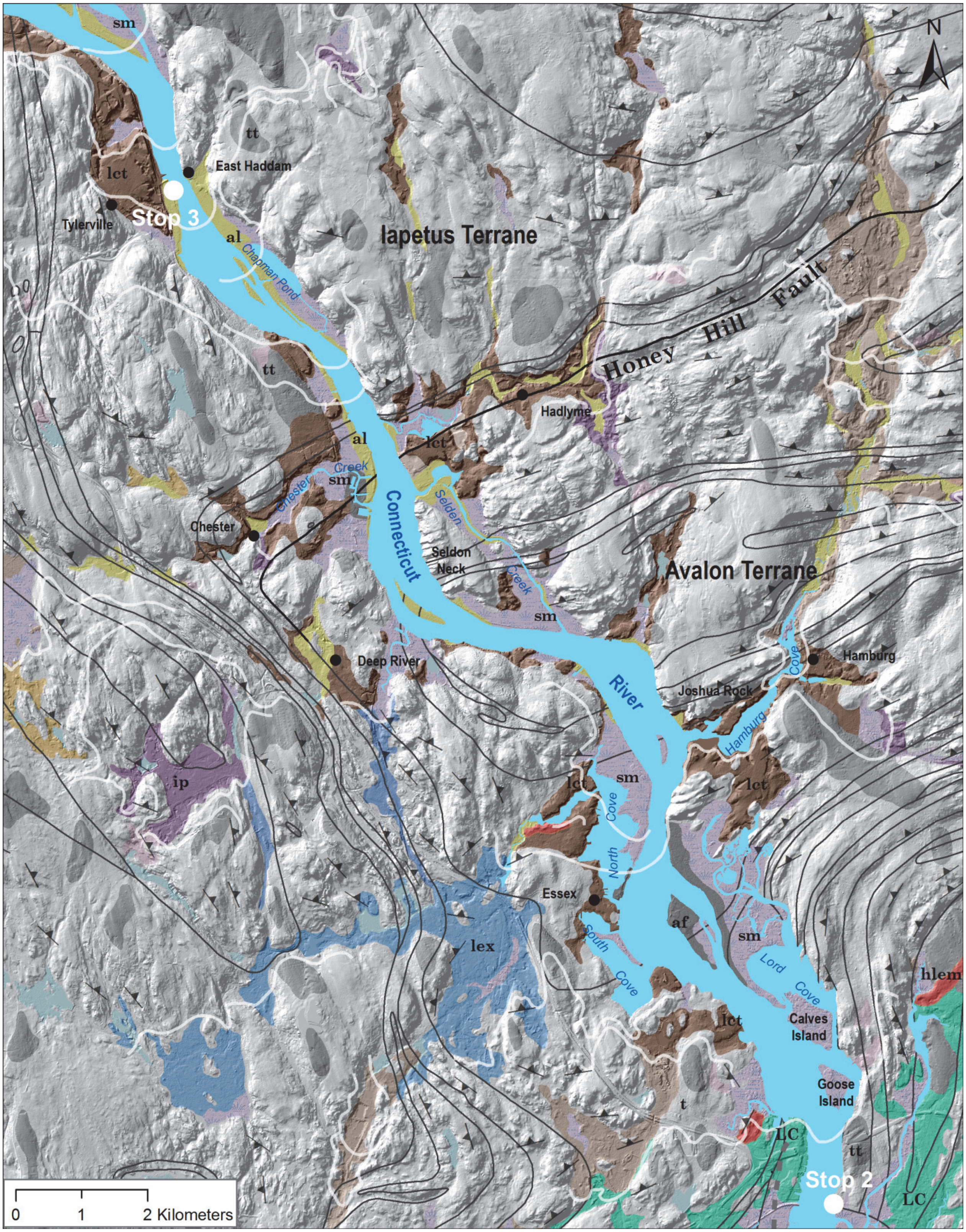


Figure 24. Quaternary geologic units along the Connecticut River from Stop 2 to Stop 3, modified from Stone and others, 2005. Heavy lines are bedrock units (Rodgers, 1985). **LC**—glacial Lake Connecticut deposits, **lct**—lower Connecticut River deltaic deposits, **lex**—glacial Lake Essex deposits, **ip**—ice-dammed pond deposits, **tt**—thick till, **sm**—salt-marsh and freshwater tidal-marsh deposits, **al**—floodplain alluvium.

Salt-marshes flank both sides of the River in many places. Salt-marsh peat was deposited as sea level has risen over the last 4,000 years; it overlies drowned floodplain alluvial deposits. The river valley is wider in the lower estuary region and the mouths of tributary valleys like Eightmile River in Hadlyme, Falls River in Essex, Deep River Stream in Deep River, and Pattaconk Brook in Chester have been flooded by the rising estuary to produce freshwater tidal coves like, Hamburg Cove, North Cove, Pratt Cove, and Chester Creek (fig. 24).

The glacial Lake Connecticut delta at the Baldwin Bridge in Old Saybrook most likely completely blocked the valley, so when the ice margin retreated northward from there ponding continued. Ice-marginal deltas of unit lct are not contiguous in this broader section of the valley; some deltas were built into ponded water in the valley from tributary valleys, as at Chester and Hadlyme. Although these deltaic deposits are more localized than farther north, delta surfaces rise steadily northward from about 10 m at Essex, to 17 m at Deep River, and 20 m at Chester.

The Honey Hill fault zone, which is the terrane boundary between Iapetan and Avalonian rocks, crosses the river between Hadlyme and Chester (fig. 24). Downstream (southeast) from there, Avalonian bedrock ridges strike nearly east-west or northeast-southwest, perpendicular to the valley, and strike-parallel tributary valleys locally enter the main valley at places like Hamburg Cove and Chester Creek. Outcrops of resistant bedrock with near vertical foliation can be seen at Joshua Rock and Seldon Neck. North of the fault, bedrock strikes generally NNW and the structure is more parallel to the trend of the River.

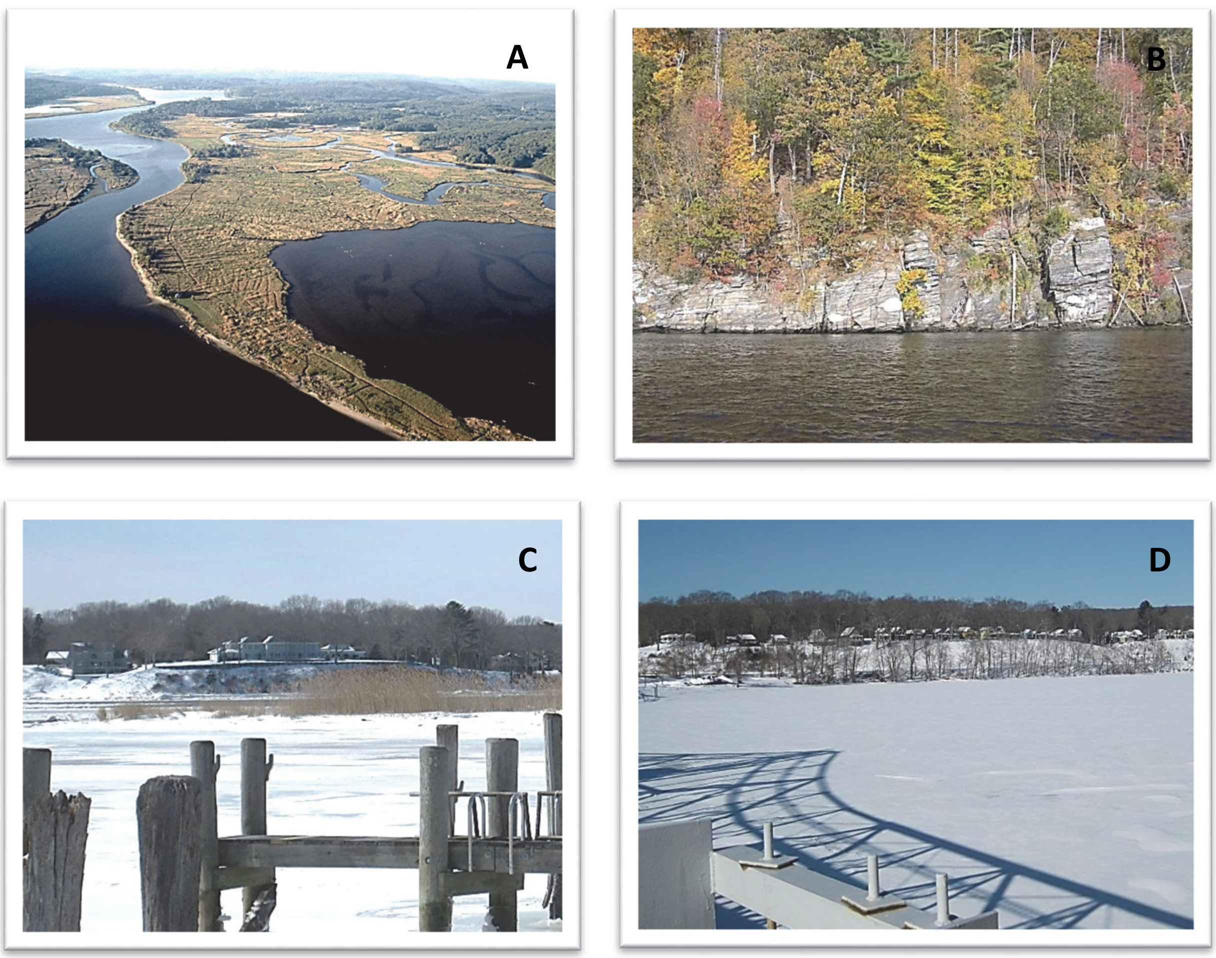


Figure 25. **A**. Looking up river, salt-marshes surrounding Lord Cove. **B**. Bedrock outcropping of boudinaged Hebron gneiss below Gillette's Castle just north of the Honey Hill fault. **C**. West view from the Hadlyme Ferry dock across River to Chester—houses sit on 20-m deltaic terrace. **D**. West view from East Haddam Bridge across River to Haddam—houses sit on 25-m deltaic terrace.

Stop 3. Eagle Landing State Park. 14 Little Meadow Road, Haddam (-72.465, 41.450)

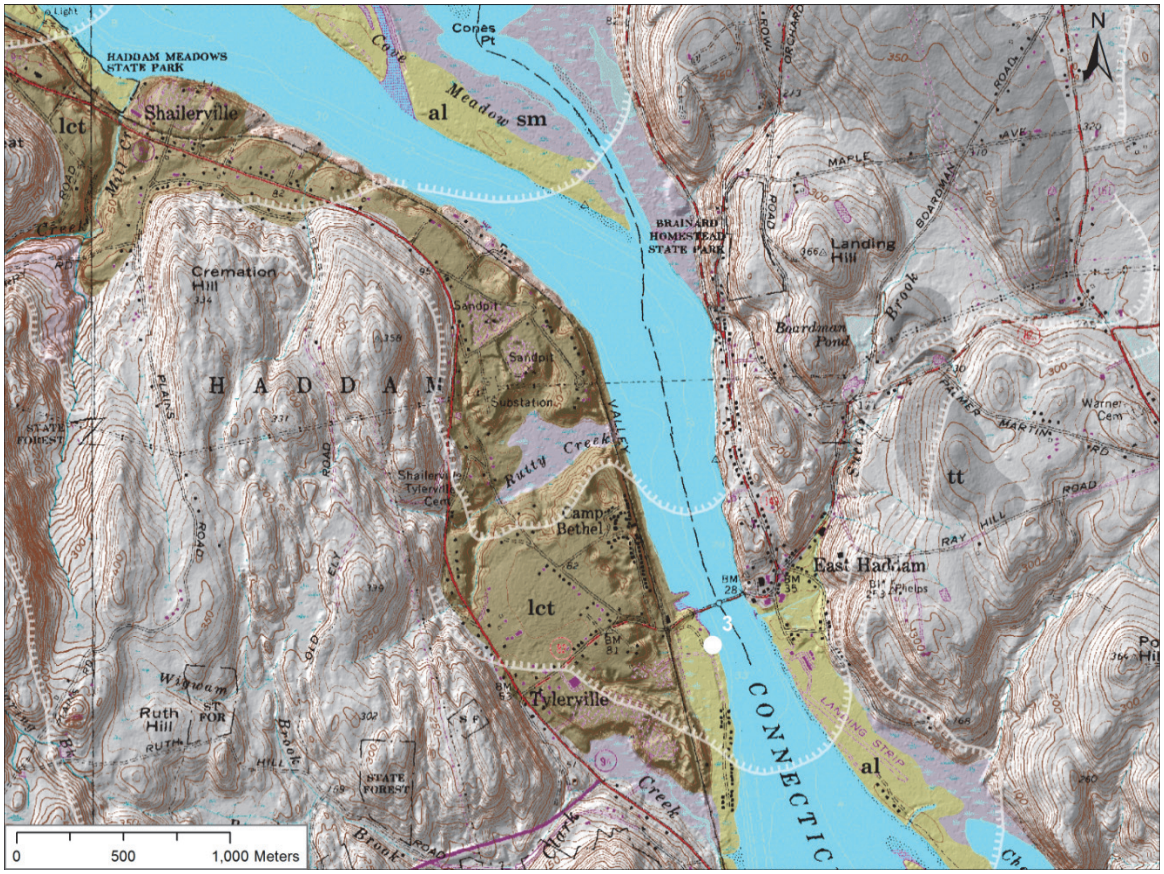


Figure 26. Quaternary geologic units in the vicinity of Stop 3 modified from Stone and others, 2005. **lct**—lower Connecticut River deltaic deposits, **al**—floodplain alluvium, **sm**—salt-marsh and estuarine deposits. Ticked white lines are retreatal ice-margin positions.

We disembark the RiverQuest onto the modern floodplain surface of the Connecticut River (fig. 26). The floodplain is near river level at <3 m altitude. Above the floodplain to the west is a flat-topped terrace surface at 25 m altitude (82 ft on topo map). From here for about 20 km upstream, continuous successive deltaic deposits are present on one or both sides of the River. On the east side of the valley, there are stretches where there is no terrace today, but deposits formerly extended completely across the valley from side to side. Small patches of gravel and sand can be found the east side of the River at and below 25-34 m but are too small to be mapped at 1:24,000-scale. Delta surfaces rise from 25 m here at the E. Haddam bridge to about 34 m at the southern boundary of Middletown—a slope of 1.1 m/km, most of which can be accounted for by postglacial tilting.

Enroute to Stop 4. We turn left out of the Eagle Landing parking area, cross the RR tracks and bear right on Camp Bethel Rd; climbing up the hill, we ascend to the top of an ice-marginal deltaic morphosequence within the **lct** deposits. The terrace surface here is at 25-m altitude (82 ft at road intersection on topographic map, fig. 26). To the east of the road, 10-15-m deep kettles mark the delta surface. With permission we can turn into Camp Bethel and take a look at them. The non-collapsed delta surface reaches 27.5 m (90 ft) at its highest point. Continuing north we descend the ice-marginal slope (large cobbles and boulders can be seen on the slope) into the valley of Rutty Creek. We turn north again on Rt. 154 and continue onto the surface of the next successive delta north of Rutty Creek; the surface of this delta rises northward from about 23 m at its distal end (dammed behind the previous one) to 26 m at the north end. A topset-foreset contact exposed in a former sand pit east of the road is at approximately 24 m in altitude (fig. 27). As we continue on Rt. 154 for about 2 km, we drive along the contact between the steep, locally till-covered bedrock slope on the left (west) and the relatively flat **lct** terrace surface to the right (east). At Shailerville, we descend off the **lct** terrace surface down to Connecticut River level as we cross the erosional valley of Mill Creek, then rise again onto the **lct** deltaic terrace that reaches 31-m altitude here. Haddam Meadows State Park to the right occupies an extensive floodplain surface that has not yet been inundated by the rising Connecticut River estuary.

Figure 27. Glaciodeltaic topset and foreset strata formerly exposed in sand pit excated in the Lower Connecticut River deposits (unit **lct**) in Tylerville. Oblique view shows 3 m of sand and gravel topset beds disconformably overlying 2 m of sand and pebble gravel foreset beds.

Stop 4. Haddam Meadows State Park. Rt. 154 Haddam (-72.507, 41.480)

Haddam Meadows State Park (fig. 28) occupies the Connecticut River floodplain at 3-4 m above present sea level. To the east across the River we can see the former site of the decommissioned CT Yankee nuclear power plant; spent fuel rods are now stored above ground on a 2-ft thick concrete pad that sits in a glacial meltwater channel (fig. 30), discovered by R.J. Lougee, who placed more significance on it than we do today. A telegram to Professor Goldthwait datelined East Haddam, Connecticut, December 11, 1948 reads—*"No longer speculate Hitchcock outlet. Waterfall site Niagaran height discovered."* Lougee never considered the New Britain spillway to be the outlet for Lake Hitchcock.

Figure 28. East view from Haddam Meadows

Looking westward with leaves off (fig. 29) we see the higher terrace of glaciodeltaic deposits at 29 m (95 ft) altitude (fig. 30). Here on the floodplain, alluvial overbank sand deposits are about 2-3 m thick (fig. 31). The modern floodplain sediment overlies about 3 m of coarser sand and gravel interpreted as stream terrace deposits laid down by meltwater exiting the New Britain spillway. The base of this terrace is at -3 m in altitude and is on grade with the -22-m base of the same channel at Stop 2. The channel altitude here with depression was -107 m, only 1 m higher than at the Baldwin Bridge. East of the floodplain, beneath the River, a deeper incision into glacial lake deposits to about -15 m is filled with estuarine sediment. This downcutting was accomplished as glacio-isostatic uplift occurred and sea level fell. As the rate of eustatic sea-level rise overtook the rate of uplift, marine transgression began and continues today in the Connecticut River valley.

Figure 29. West view from Haddam Meadows. House sits on terrace of glaciodeltaic deposits at 29 m in mid-background.

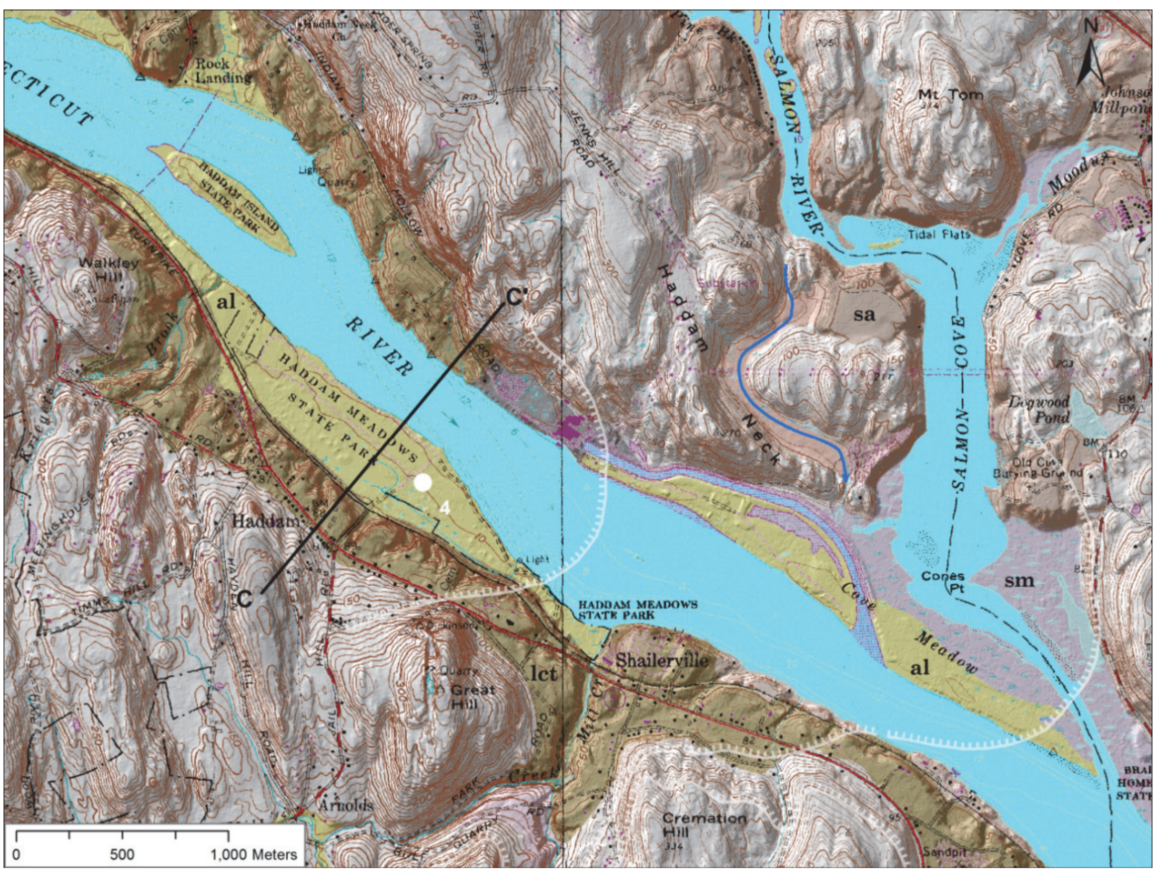


Figure 30. Geologic units in the vicinity of Stop 4 modified from Stone and others, 2005. **lct**—lower Connecticut River deltaic deposits, **al**—floodplain alluvium, **sm**—salt-marsh and estuarine deposits. Ticked white lines are retreatal ice-margin positions. Blue arrow east of Haddam Neck shows meltwater channel at the end of which is Lougee's "waterfall site Niagran height".

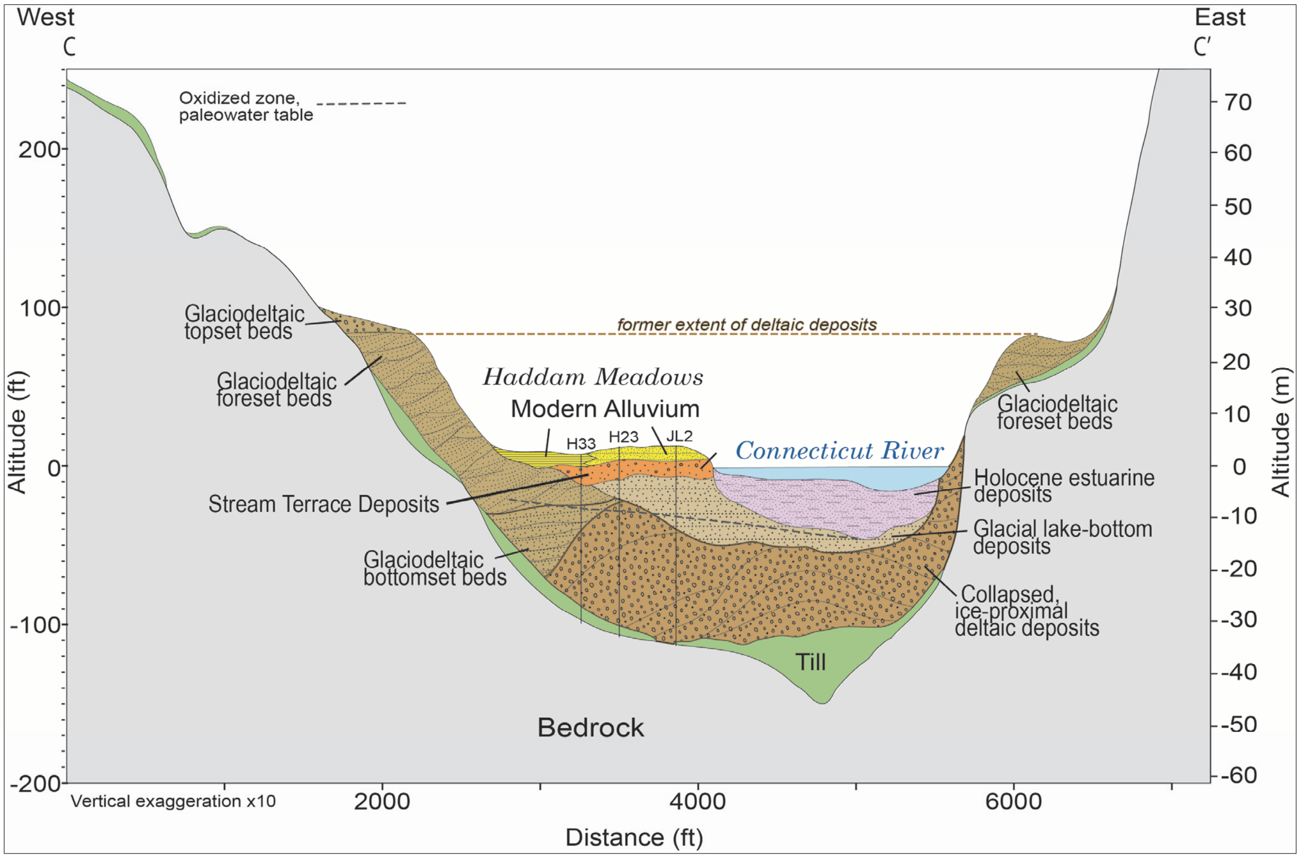


Figure 31. Cross-section C-C′ at Haddam Meadows constructed using well logs, continuous core holes, GPR reflection profiles, marine seismic reflection profiles, and regional geologic models.

Haddam Meadows has been used as a geophysical equipment testing ground by the USGS Office of Groundwater, Branch of Geophysics for many years. Therefore, the geology beneath Haddam Meadows has been thoroughly defined using many types of surface and borehole geophysical data and test borings. We will show some of the geophysical records on the fieldtrip.

Stop 5. View of "The Straits" of the Connecticut River, River Road, Middletown (-72.597, 41.557).

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 in a gorge cut through ledges of Collins Hill Formation (fig. 32). Bedrock beneath the river in the gorge is relatively shallow since this route is actually a postglacial drainage diversion—the deep bedrock valley is actually beneath the Jobs Pond valley. Further east along River Road looking across the River, we can see bedrock outcrops drop off and disappear beneath the glacial meltwater sediments where the buried ancestral Connecticut River valley joins the present course. East of the Straits, the River cuts through the 45-m surface of glacial meltwater deposits in the Jobs Pond valley. These deposits were continuous from side to side in the valley as the ice margin retreated northward here. Before incision, this sediment provided the dam that impounded glacial Lake Middletown in the basin to the west.

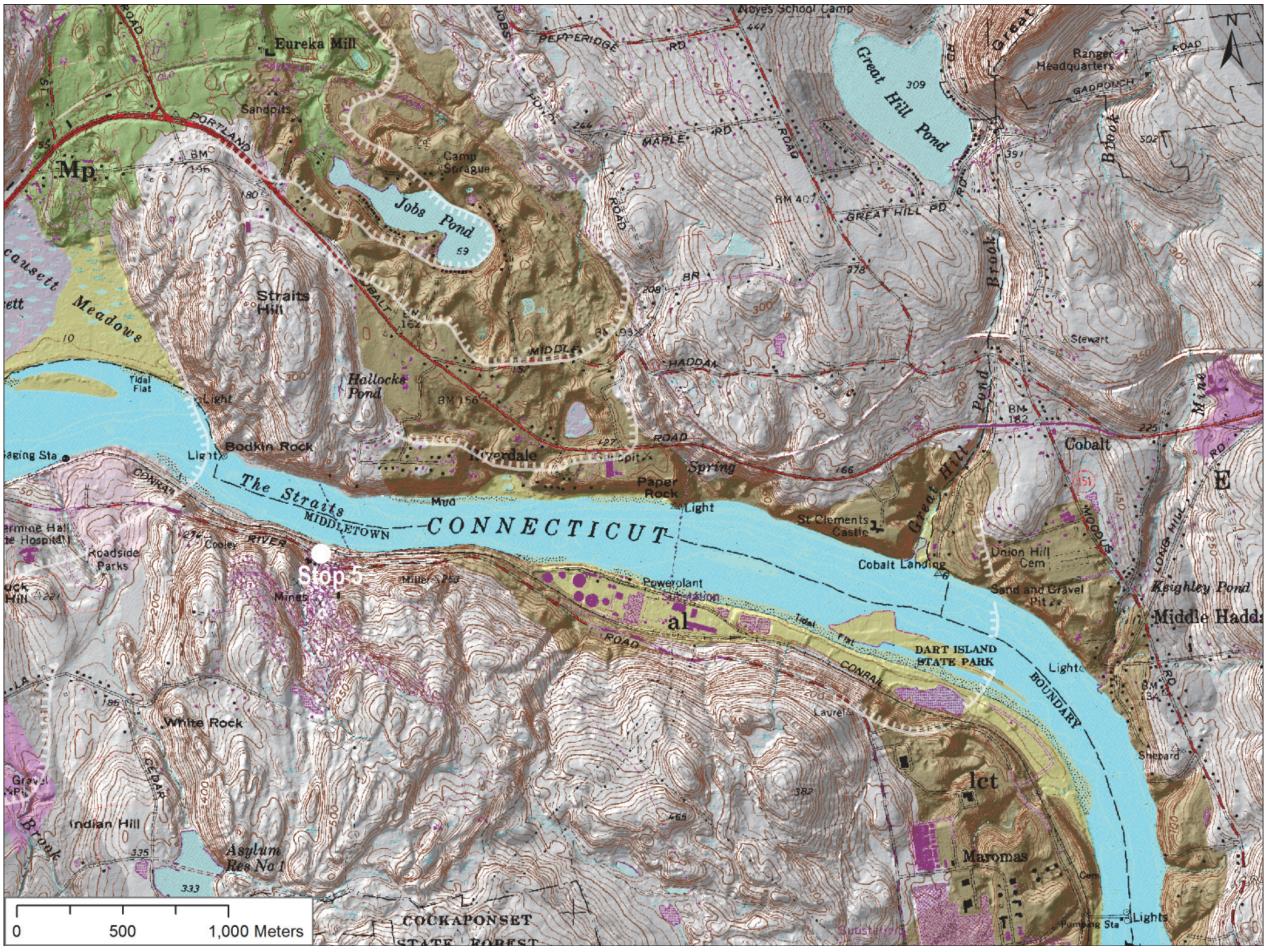


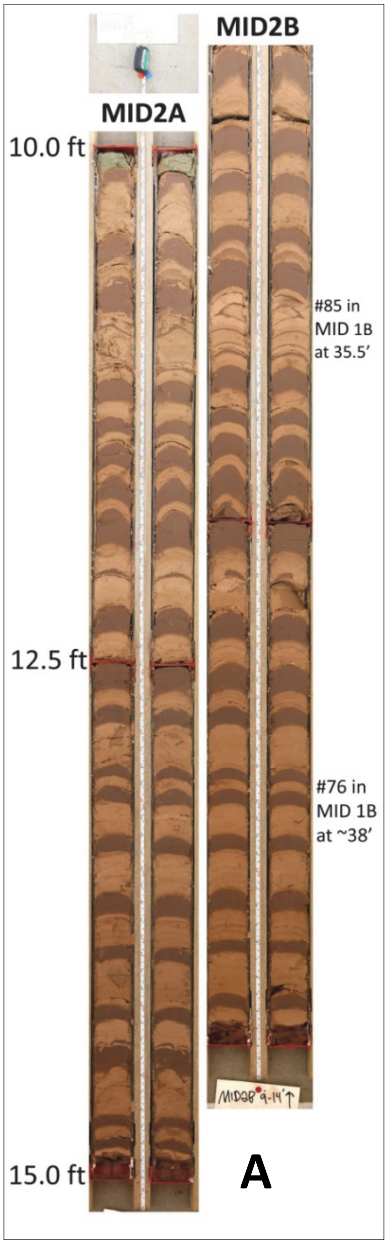
Figure 32. Quaternary geologic units in the vicinity of Stop 5 modified from Stone and others, 2005. **Mp**—glacial Lake Middletown deposits; **lct**—lower Connecticut River deltaic deposits, **al**—floodplain alluvium. Ticked white lines are retreatal ice-margin positions.

From the view point, proceed about 1800 ft west on River Road (fig. 32) to examine an exposure of old till in a bank on the south side of the road. 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°. 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 Mesozoic rock derived 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, or upper till. Larger boulders are present in this upper unit. The exposure face needs to be cleared off and examined more closely.

Stop 6. Lake-bottom deposits of glacial Lake Middletown, Lawrence School, off Mile Lane, Middletown (-72.681, 41.589)

Figure 33. Quaternary geologic units in the vicinity of Stop 6 (modified from Stone and others, 2005). **Mb**—lake-bottom deposits; **Mc**—delta deposits, **st**—stream terrace deposits, **al**—floodplain alluvium, **sm**—tidal-marsh deposits. Numbered circles are Antevs 1928 localities from which his Berlin and Newfield Series curves were generated.

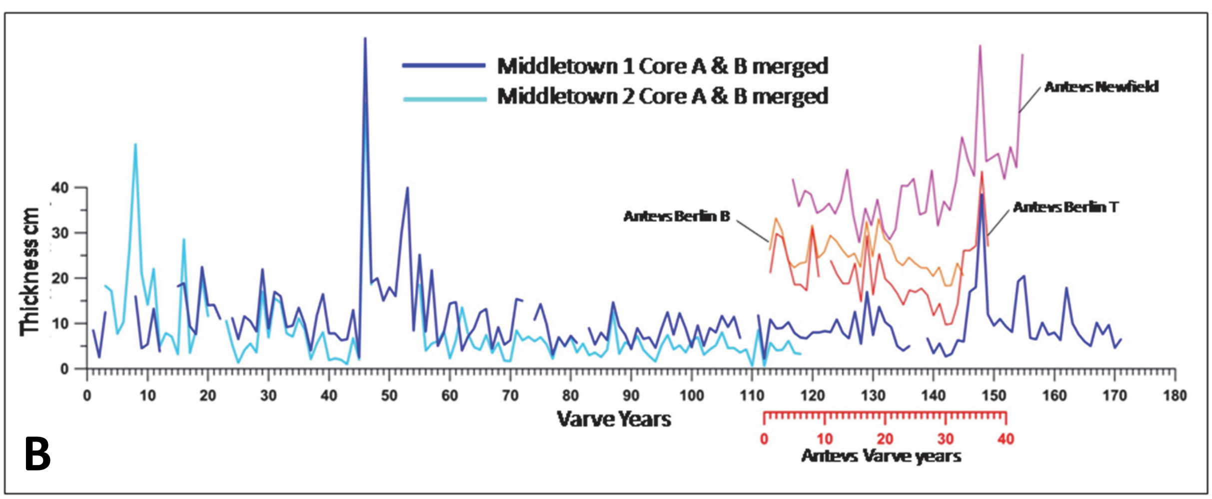
Lawrence School sits slightly higher than the lake-bottom surface on a small drumlin that was not covered by lake-bottom sediments although it is well below the approximate 37-m (120-ft) lake level of glacial Lake Middletown (fig.33). Varved lake sediments were presumably deposited by underflow currents descending from delta fronts and accumulated only in the deepest parts of the lake. Varved lake-clay surfaces are preserved at 11-14 m (35-45 ft) in altitude where they have not been postglacially eroded by Swamp Brook and its tributaries and the Masttabesset River. The lake clay is thickest near the Mattabesset River. 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 (**st** on fig 33) constructed along the Mattabessett River by water exiting the New Britain spillway from glacial Lake Hitchcock to the northwest. The terrace surface is at 6-7 m (20-25 ft) altitude. 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. 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. We will bring sections of the core discussed below along on the fieldtrip.



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. 33, 34A). 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. 34B). 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. 34B) and add 16 additional years to the chronology.

Figure 34A 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). The lake clay section is 75 ft thick at the MID 1 site and only 39 ft thick at the MID 2 site, but the same lower 119 varves are present at both sites and are thicker in the deeper parts of the lake. MID1 core curve shows 62 more years than MID2 because the upper part was badly deformed and varves could not be measured. 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 also has been subjected to the deformation that formed the rimmed depressions interpreted to be pingo or lithalsa scars.

Figure 34. **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.



Numerous small circular to subcircular vernal pools with subtle raised rims are present in the wooded area just northwest of the Lawrence School (fig. 35). 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. Similar to the lake-bottom surfaces of glacial Lake Hitchcock and other glacial lakes in southern New England, those of glacial Lake Middletown exhibit clusters of small rimmed depressions that have been interpreted as pingo scars (Stone and Ashley, 1992). These features may actually be remnants of lithalsas that formed on the drained lake bed through the process of cryosuction in a discontinuous permafrost zone. 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.

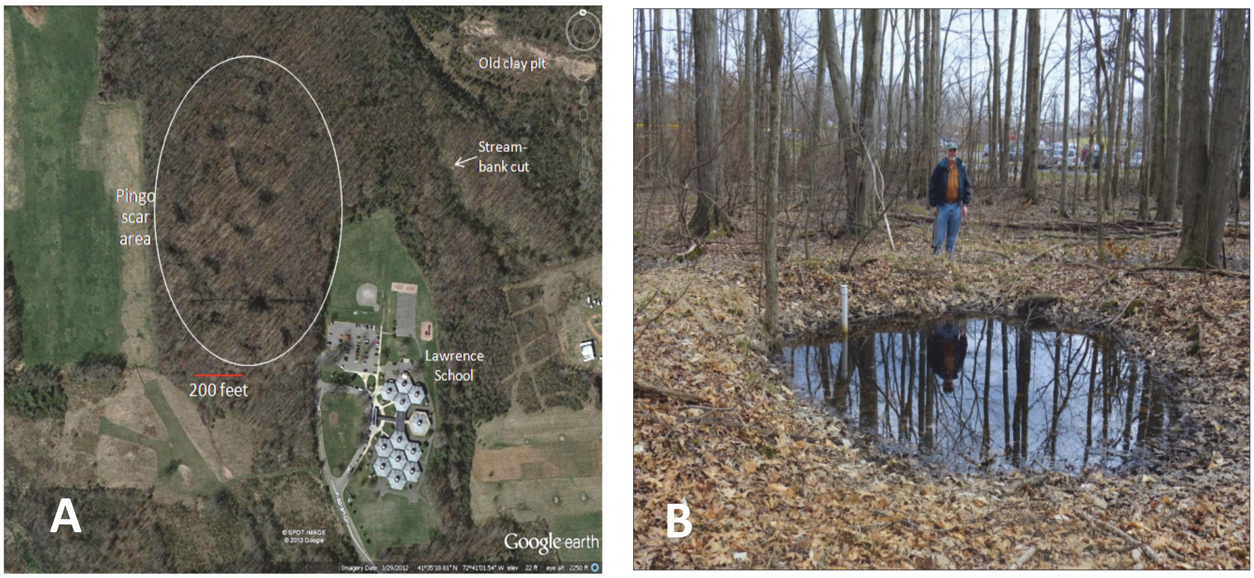


Figure 35. **A.** Google image March 2012; dark, circular areas just northwest of school are small vernal pools in bottoms of pingo/lithalsa-scar depressions. **B.** Photo of depression in the woods just west of Lawrence School; note subtle rim.

Day 2

Stop 7. Dividend Pond Open Space and Rocky Hill dam view, Old Forge Road, Rocky Hill (-72.634, 41.646)

We access the old "Mustard Bowl" gravel pit in the Rocky Hill dam (fig. 36) visited on many previous fieldtrips via a new route; first, a scenic waterfall cascading over bedrock and the contact between Holyoke Basalt above and East Berlin Formation below. The east-west trending bedrock ridge (**Jha**, fig. 37) is part of the southern limb of a broad fold in the layered Mesozoic bedrock called the Rocky Hill anticline (Rodgers, 1985). Because of this folded bedrock structure producing east-west striking resistant ridges of basalt, the Connecticut River valley is significantly narrower in the Rocky Hill – Glastonbury vicinity; and it is here that the Rocky Hill dam for glacial Lake Hitchcock was emplaced. We will follow the trail from the waterfall (fig. 36B) up along the rim of the old gravel pit to its highest point where we have a view (leaves permitting) across the River to the surface of the Rocky Hill dam on the east side. There are currently no fresh faces in the gravel pit, but horizontally bedded gravel topset beds overlying sandy dipping foreset beds can be seen in the 1986 photo (fig. 36A).

The pit is excavated into deposits of unit **Mdb** which is an ice-marginal deltaic morphosequence that filled a small lake immediately behind the unit **Mc** delta (fig. 37). Water spilled out through the Dividend Brook spillway (**dbs**) across the **Mc** delta surface. If Lake Middletown had not still existed at the end of this spillway, preventing further incision, the outlet for Lake Hitchcock might have developed here instead of at New Britain.

Figure 36. **A.** Mustard Bowl Pit excavation as it was in 1986. 3-4 m gravel topset beds overlie 25 m southwest dipping sandy foreset beds. **B.** Dividend Pond falls over Hampden Basalt (**Jha**) and East Berlin Formation (**Jeb**).

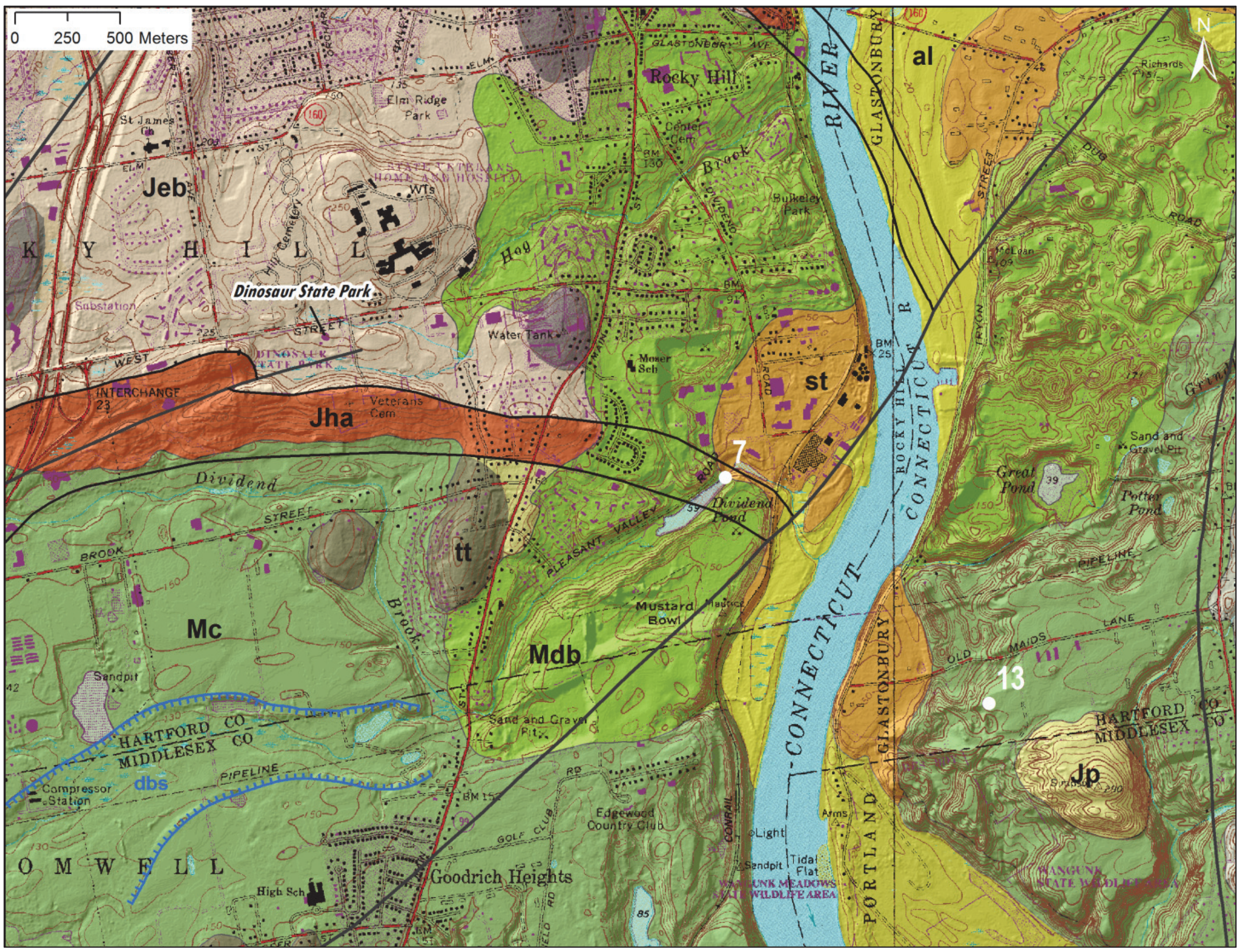


Figure 37. Bedrock and Quaternary geologic units in the vicinity of Stop 7 and 13 modified from Rodgers, 1985; Stone and others, 2005). **Jha**—Hampden Basalt, **Jeb**—East Berlin Formation, **Jp**—Portland Formation, **Mc**—Cromwell deltaic deposits of glacial Lake Middletown, **Mdb**—Divendend Brook deltaic deposits, **st**—Stream-terrace deposits (15-m terrace), **al**—Floodplain alluvium, **tt**—Thick till deposits (drumlins).

Stop 8. New Britain Spillway, end of Stamm Road, Newington (-72.748, 41.679)

The New Britain channel was first suggested as the spillway outlet for glacial Lake Hitchcock by R.F. Flint (1933). Morphology of the spillway is best seen on older topographic maps (fig. 38A) because today the area is highly modified by modern development. Water spilling from glacial Lake Hitchcock flowed through the New Britain spillway during the high levels of the lake from 17.9 cal ka until 17.2 ka and during its longer-lived stable phase from 17.2 until 15.6 cal 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 spillway is approximately 4 km in length and 250-300 m in width. It is a swampy, underfit channel not occupied by any major stream, and was cut into Lake Middletown deposits, till, and incompetent shaley bedrock from an altitude of about 35 m down to its present 17-m level. An 8-m deep water column in the channel produced the 25-m stable level. Previous studies of glacial Lake Hitchcock have considered the New Britain spillway as the only base-level control for the lake. But incision of this channel could only have occurred as base levels to the south lowered; while the land was glacio-isostatically depressed, there was no fluvial gradient south of the spillway at its stable 25-m level (fig. 14), and sea level was rising. Therefore, Lake Hitchcock could not lower further until glacial rebound began.

Figure 38.**A.** Topography of the New Britain Spillway. Lidar-derived hillshade overlain by U.S. Geological Survey topographic maps, scale 1:31,680; New Britain 1946; Hartford South 1944. Shaded blue area is spillway.

B. Till exposed in 1982 in the east wall of the New Britain spillway about 500 m south of New Britain Ave. Elizabeth Haley London for scale.

Stops 9-11 are in the Farmington River delta complex, which consists of deposits built into 3 different levels of glacial Lake Hitchcock (see fig. 4 and discussion in text) and higher ice-hole deposits likely controlled by glacial Lake Middletown water levels.

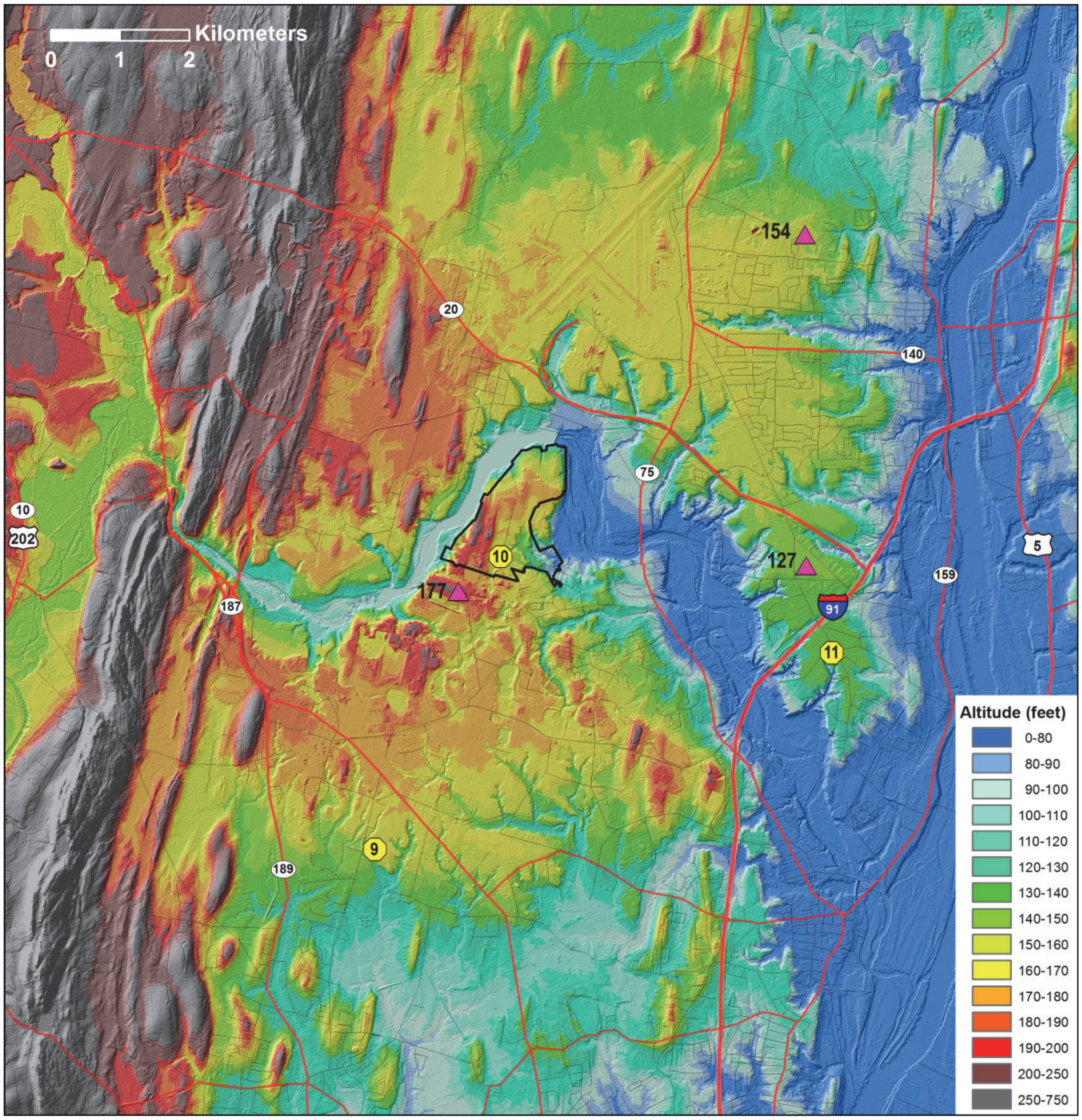


Figure 39. Farmington River delta complex displayed as a lidar-derived digital elevation model (DEM) (1-ft vertical resolution). Heavy black line is Northwest Park boundary. Pink triangles are topset-foreset contacts measured in high-level, stable-level, and post-stable level deltas; altitudes in feet. Yellow circles are STOP numbers.

Figure 39 displays the delta complex as a digital elevation model colored for 10-ft ranges in altitude. The high-level delta has surface altitudes of 170-190 ft and shows as orange colors getting lighter (lower in altitude to the southwest); topset-foreset contact at 177 ft. The stable-level delta has surface altitudes in the 160-170-ft range and shows as yellow getting lighter to the northeast; topset-foreset contact at 154 ft. The post-stable delta has surface altitudes at 130-140-ft range (green) and slopes to the southeast; topset-foreset contact at 127 ft. However, because the delta complex covers about 8 km of north-south extent, there is considerable differential glacio-isostatic tilt incorporated in its present position. Figure 39A corrects for the differential tilt by subtracting the estimated total depression surface (fig 15) from the above DEM and converting to meters. The resulting DEM (fig. 39A) depicts the parts of the delta complex at altitudes as they were before glacio-isostatic rebound occurred—the stable level delta was at -100 m in altitude (see fig. 16).

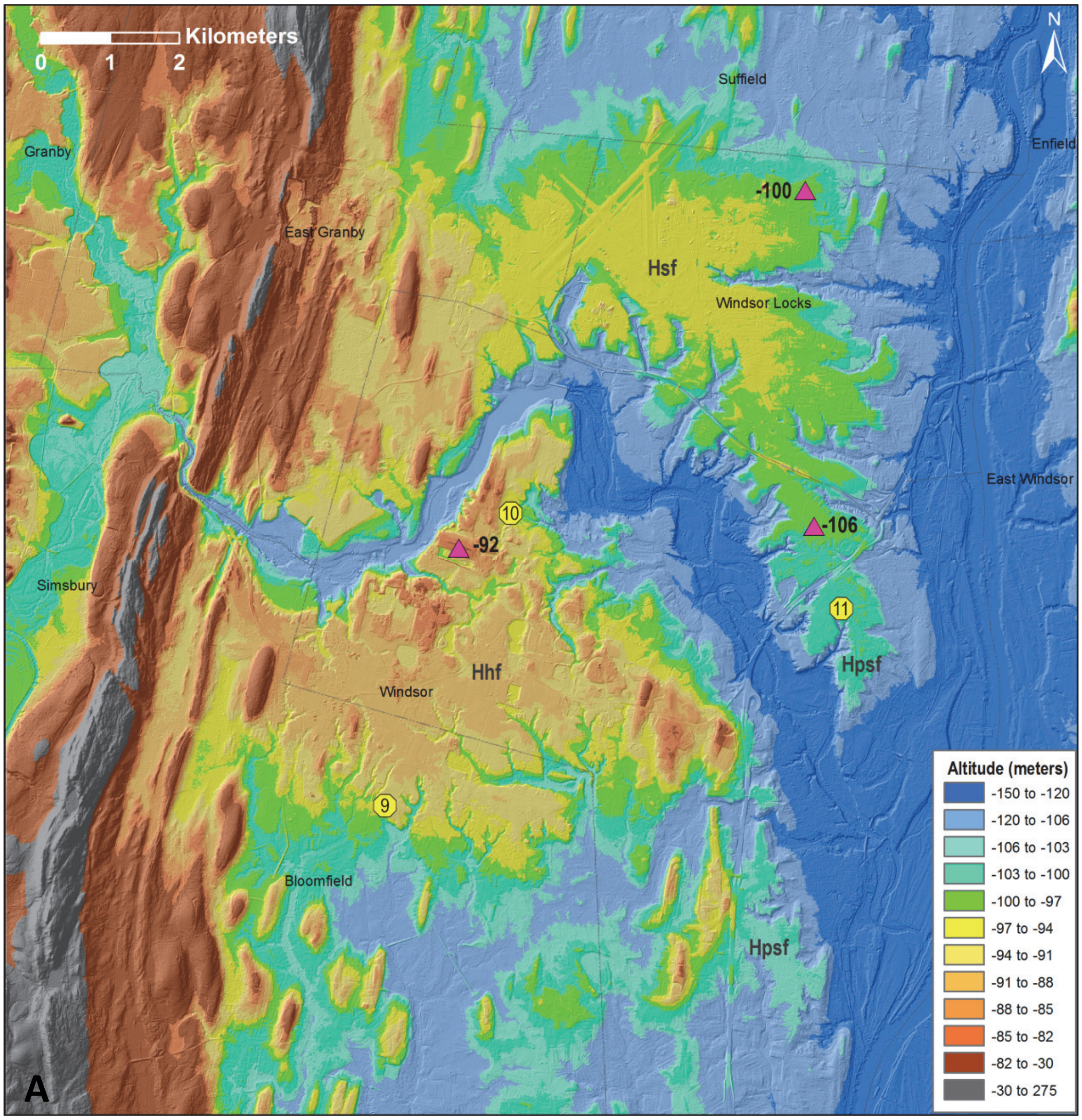
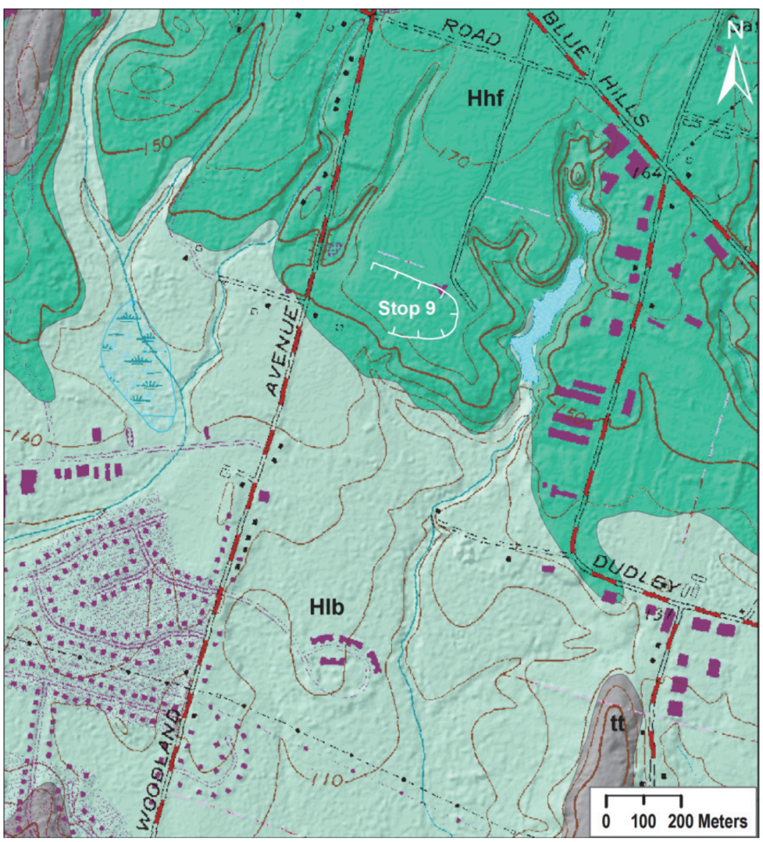


Figure 39**A**. Farmington River delta complex displayed as a DEM after adjusting for total depression and converting to meters. Pink triangles are topset-foreset contacts measured in high-level, stable-level, and post-stable level deltas. Altitudes are adjusted for glacio-isostatic depression (converted to meters)

Stop 9. Woodland Road excavation, 390 Woodland Road, Bloomfield (-72.726, 41.862)

The excavation (fig. 40, 41) is located in the distal most part of the Farmington River high level delta (**Hhf**, fig. 41). This delta was built southerly into the lake from an ice-margin position 4 km to the north of here. Less than a meter of fine-grained topset beds are exposed in the northern and eastern parts of the face, although we can argue whether the sand unit that contains the swallow burrows is a fluvial bed or a foreset bed. About 2 m of sandy foreset beds, dipping southeasterly are exposed in the southern pit face. The topset-foreset contact is not currently exposed but is estimated at 49-50 m (160-163 ft). When adjusted for differential glacio-isostatic tilting, this altitude represents a 35-m level at the New Britain spillway.

Figure 40. North face in the Woodland Road excavation.



The lake-bottom surface (**Hlb**, fig. 41) lies immediately south of here at an altitude of 41-44 m (135-145 ft). At times, gray clay is exposed in the floor of the excavation.

Enroute to Stop 10: Leaving the excavation we drive north on Woodland Road, heading directly up the delta surface plain. We turn left on Blue Hills Ave., then right on Day Hill Rd., skirting the edge of slightly older ice-contact deltaic deposits (unit **Mg**, fig. 4) on the left. Turning left on Prospect Hill Road, we quickly begin to see the flat delta plain become more rolling, and several large kettles (Great Pond and Silver Birch Pond) deform the plain. We turn left at the circle onto Lang Road and drive down the ice-marginal slope to Northwest Park.

Figure 41. Geologic units in the vicinity of Stop 9 modified from Stone and others, 2005. **Hhf**—High-level Farmington delta deposits, **Hlb**—Lake-bottom deposits, **tt**—Thick till deposits (drumlins).

Stop 10. Northwest Park, 145 Lang Road, Windsor (-72.704, 41.899). **Lunch Stop**

Northwest Park is a 473-acre Municipal Park owned and operated by the Town of Windsor; what was once tobacco farmland has been converted to biologically diverse forests and fields embedded with streams and wetlands. SSW trending hills in the western part of the park are underlain by glacial till (drumlins) and meltwater deposits of glacial Lake Hitchcock surround the hills. We will have lunch near the Nature Center, then walk to the south end of the Park to look at ice-marginal deposits of the high level delta. Then we will hike the purple trail (fig. 42) to examine and discuss the constructional or erosional nature of the topography along the trail, and the validity of its current map unit designation—**Hlb** (fig. 42).

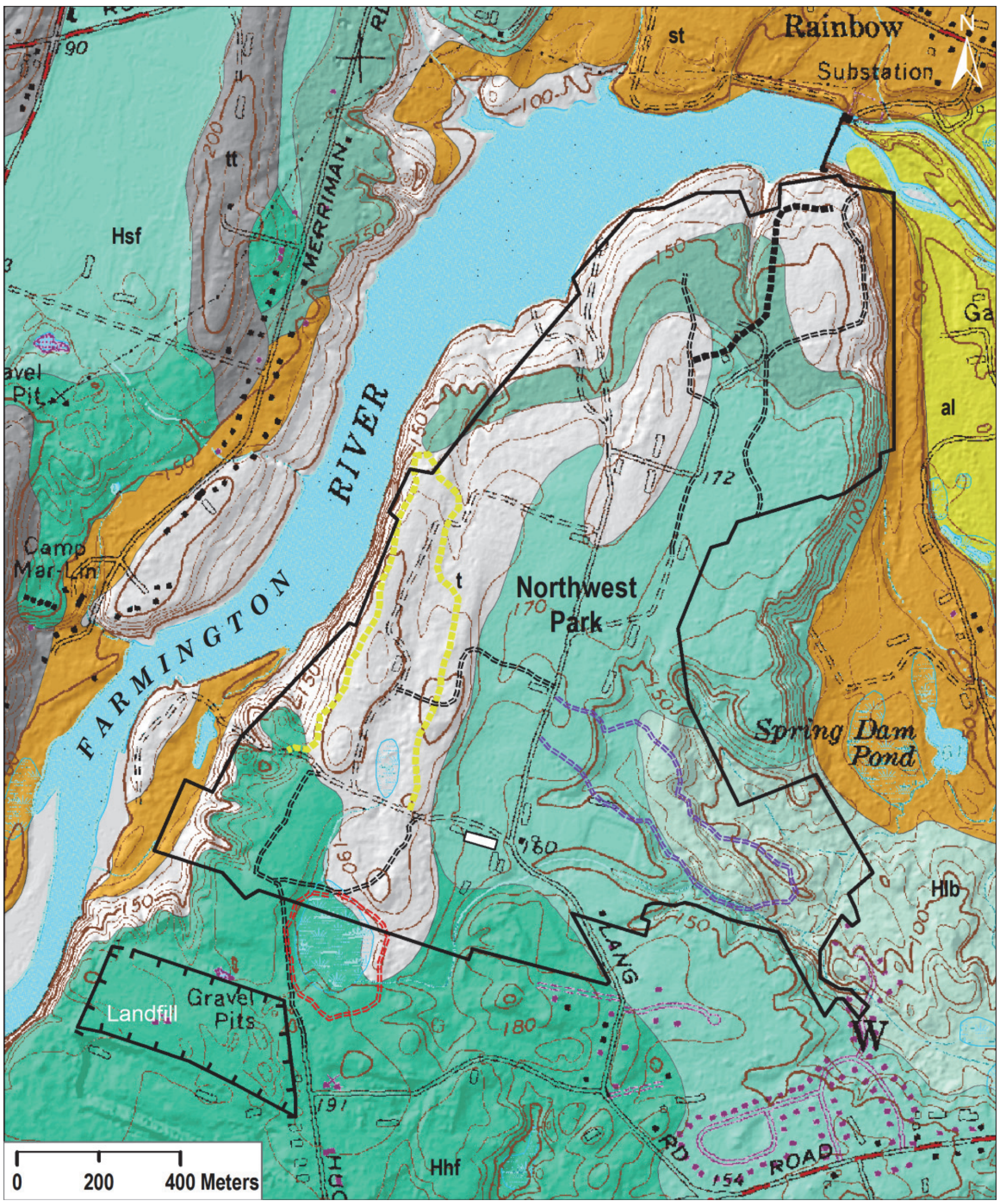


Figure 42. Quaternary geologic units in the vicinity of Stop 10 modified from Stone and others, 2005. **Hhf**—high-level Farmington delta deposits, **Hsf**—stable-level Farmington delta deposits, **Hpsf**—post-stable Farmington delta deposits, **Hlb**—lake-bottom deposits, **st**—stream terrace deposits, **al**—floodplain alluvium. White rectangle—Northwest Park Nature Center.

In the past, there have been several large sand and gravel pits (fig. 43) in the ice-marginal parts of the high-level Farmington delta deposits. Today the only remaining excavation is at the Windsor-Bloomfield Landfill just south of the Park, but no active pit faces are currently available. A topset-foreset contact (fig. 43) at an altitude of 54 m (178 ft) was measured at a former exposure at the landfill. We hike from the Nature Center west over a till hill (probably a drumlin) then south into a valley that contains a kettle-hole bog (red trail on map fig. 42). Note that the ground along the trail is stonier, and redder than back at the Nature Center. The kettle is in the northernmost part of the ice-marginal delta (unit **Hhf**). Continuing south the old road leaves the Park and goes into the north side of the Landfill; we can examine red sand and gravel in small inactive scarps. The ice-marginal deposits in the delta complex contain predominantly Mesozoic-rock derived clasts (red-brown sandstone, siltstone, conglomerate, local basalt) while the more distal deposits like the stable level delta consist of metamorphic-rock derived clasts that came from the Farmington River coming out of the western highlands.

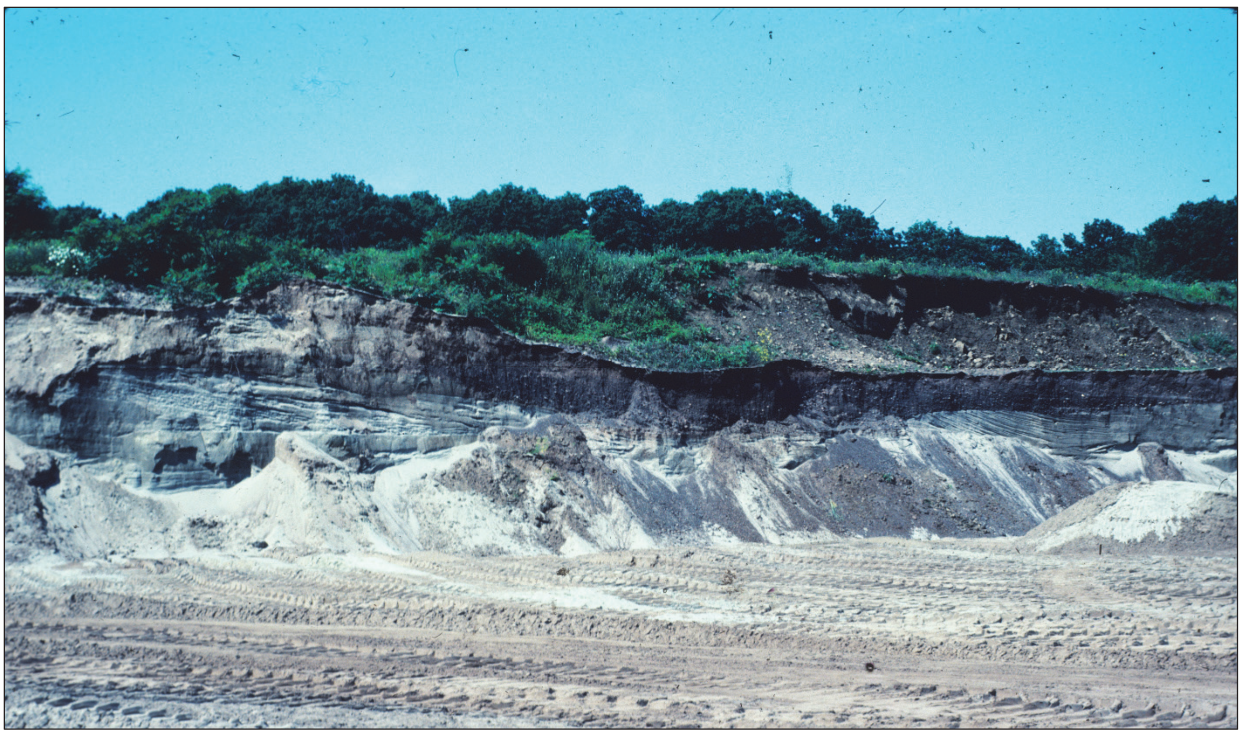


Figure 43. Former exposure of red gravel and white sand in unit **Hhf** at the Windsor-Bloomfield Landfill.

Taking the old gravel road past the historic tobacco barns to the north end of the Park, we can look across the Farmington River (dammed here forming Rainbow Pond) to the stable-level delta surface (**Hsf**, fig. 42) that stands above stream terrace deposits (**st**). This surface has been slightly incised by the fluvial feeder part of the post-stable delta immediately across the river (**Hspf**), but the extensive surface beneath Bradley International Airport is the topset plain of the stable-level delta.

Enroute to Stop 11: Exit northwest park and turn left on Prospect Hill Road at circle. Travel across the delta surface, then descend to stream terrace surface at left turn onto Poquonock Ave. (Rt. 75). Cross Farmington River and turn right on River St. Travel along floodplain at 8 m (25 ft) altitude, then ascend to stream terrace level 23-26 m (75-85 ft), and ascend again to delta level at 14 m (145 ft) in altitude. Turn right on Kennedy Road.

Stop 11. Kennedy Road Post-stable Farmington Delta, Thrall Road, Windsor (-72.646, 41.887).

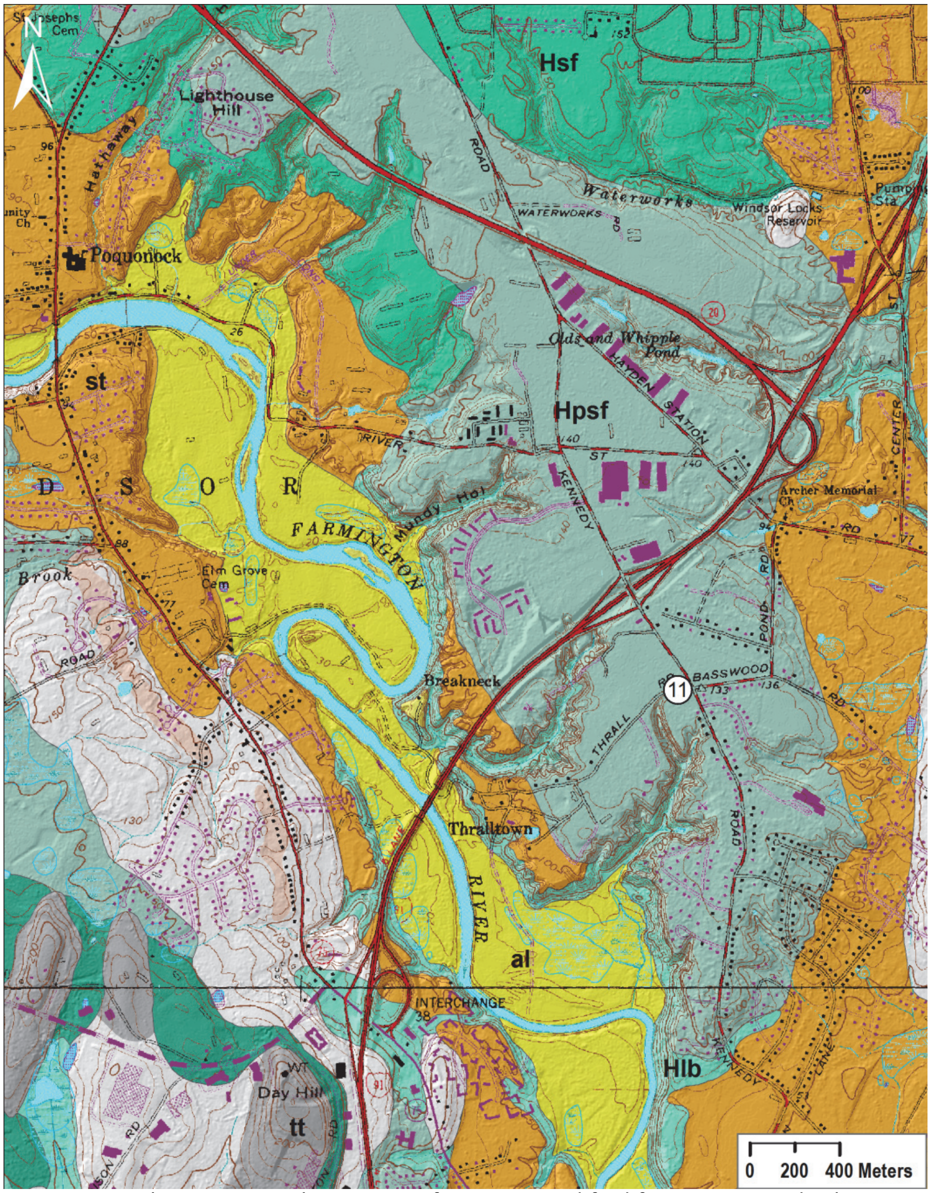


Figure 44. Quaternary geologic units in the vicinity of Stop 11 modified from Stone and others, 2005. **Hspf**—post-stable delta deposits, **Hsf**—stable delta deposits, **st**—stream terrace deposits, **al**—floodplain alluvium.

Kennedy Road traverses a lobe of the post-stable Farmington delta plain. This delta was built southeasterly into the lake by the Farmington River when the lake level lowered due to the inception of glacio-isostatic rebound and failure of the Rocky Hill dam. The river first had to cut its way through the slightly higher stable-level delta via an inset fluvial terrace (shown fig. 43 and 44 as north and west parts of unit **Hspf**). The post-stable delta is only 10-15 m thick and overlies thicker varved clay that can be seen in the lower parts of the deep gullies like Mundy Hol (fig. 44) that cut into the delta. We will stop along Thrall Road beside the fields of O.J. Thrall, Inc. (one of the largest and last remaining shade-grown tobacco farms in Connecticut; the tobacco is used to make cigar wrappers). With permission from the landowner, we may be able to examine the scarps in the gully just south of the road.

Stop 12. Matianuck Dunes Natural Area Preserve, Keney Park Road, Windsor (-72.680, 41.807). Sand dunes within the preserve are designated as a Natural Heritage Trust area because they provide rare habitat for several endangered insect species including the ghost dune tiger beetle (Cicindela lepida).

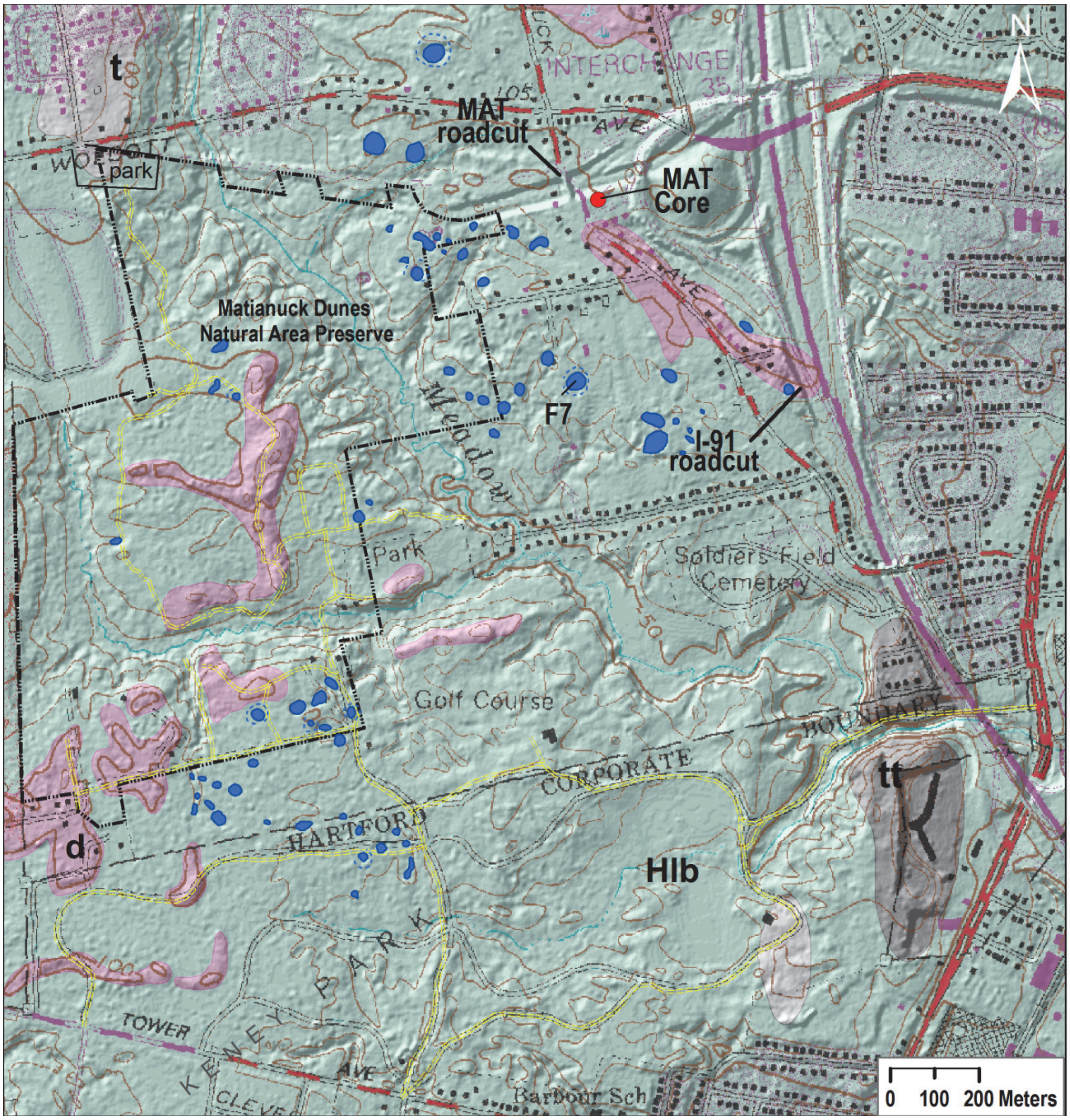


Figure 45. Quaternary geologic map units in the vicinity of Stop 12 modified from Stone and others, 2005. **Hlb**—lake-bottom deposits, **d**—inland dune deposits, **t**—thin till and shallow bedrock. White dashed line—Matianuck Dunes Preserve boundary, Gray dashed line—Preserve trails, Blue spots—rimmed depressions (lithalsa scars).

The preserve is located on the drained lake-bottom surface of glacial Lake Hitchcock. The lake-bottom surface became emergent at about 15.6 cal ka, when the high water-table, non-vegetated surface was subjected to harsh climatic conditions and discontinuous permafrost likely developed. Strong winds blew sand from exposed delta areas and constructed linear and arcuate dunes on the lake-bottom surface at the same time that permafrost mounds were forming.

Figure 46. **A.** Eolian sand dune in Matianuck Dunes Preserve**. B.** Rimmed depression in the Preserve.

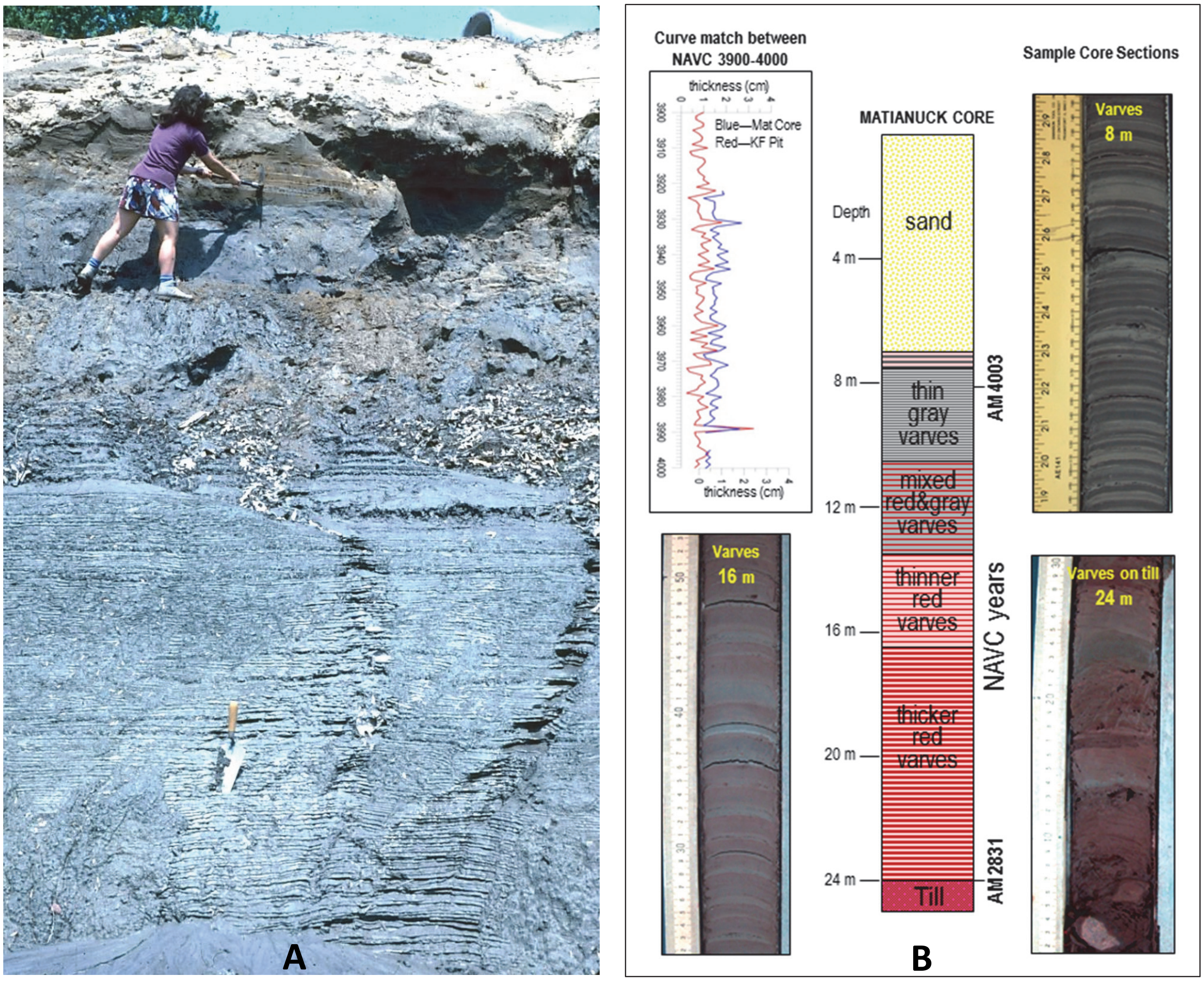
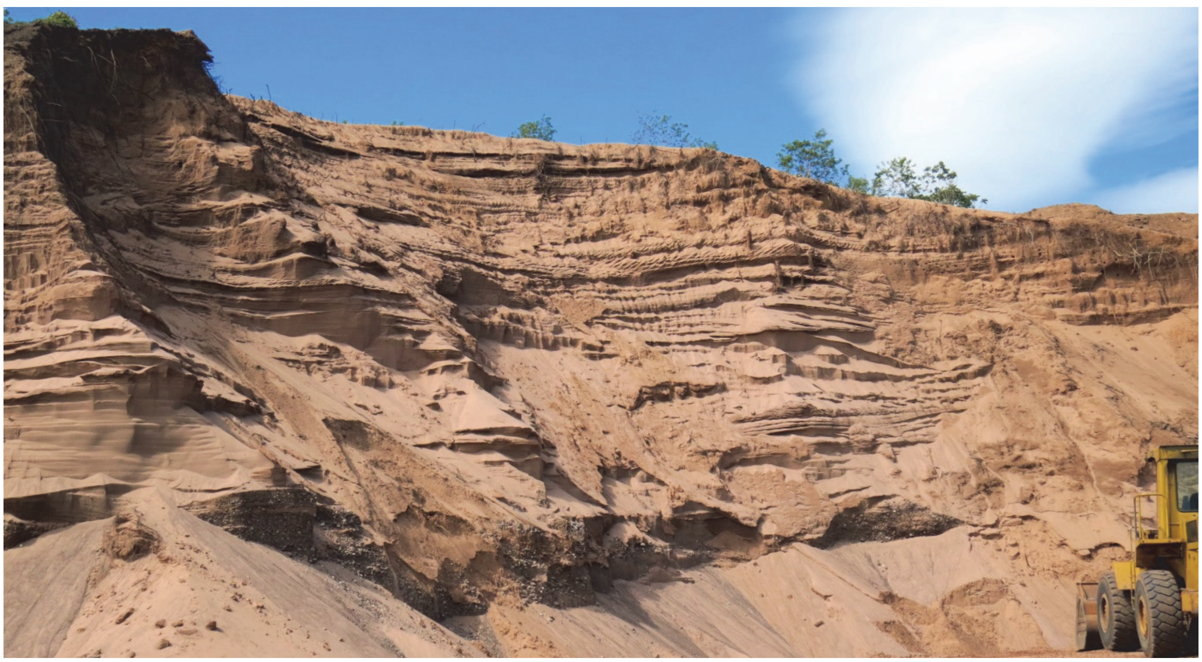


Figure 47. **A.** Matianuck roadcut exposure: section contains 1 m of eolian sand overlying 3 m of lacustrine sand overlying 3-4 m of varved clay. **B.** Matianuck Core diagram and sample core sections; upper left diagram shows curve match between upper part of core and KF plant pit between NAVC 3900-4000. Location of roadcut and core, fig 45.

Varved clay underlies eolian and/or lacustrine sand beneath the lake-bottom surface, and while no varve outcrops have been found within the Preserve, excellent exposures (fig. 45, 47A) were available not far away during construction of an access ramp for Rt. 291 in 1991-92 and for the widening of I-91 in 1988-89 (fig. 45). The Matianuck roadcut exposed eolian and lacustrine deposits (fig. 47A), and the I-91 roadcut exposed a cross-section through one of the rimmed depressions. These localities also provided a number of radiocarbon dates. From lacustrine sands in the Matianuck roadcut--dates on detrital plants (well-preserved dryas leaves) are 13,540±90 (Beta-59094, CAMS-4875) (Miller, 1995), 13,735±180 (W6397), and 14,120±90 (Beta-52711) (Stone and Ashley, 1992). From the I-91 roadcut, a date of 14,330±430 (Beta-35211) on detrital wood fragments in deformed lacustrine silt below rimmed depression fill of eolian sand and peat. Dates from the basal peat fill are 12,090±110 (AMS date on spruce needle) and 11,890±110 (Beta-34820) (Stone and Ashley, 1992). Two radiocarbon dates from lower peat fill in lithalsa scar F7 just east of the Preserve (fig. 45) are 12,630±240 (Beta-46514) and 12,320±110 (Beta-46513) (Robert Gelinas, Ebasco Services, Inc., written commun., 1990). Collectively these dates indicate that from 16.9-16.1 cal ka, Lake Hitchcock was still in existence in Connecticut. Then the lake drained (at 15.6 cal ka as indicated by varve data and deltas in Massachusetts), and by 14.2 cal ka, the rimmed depressions had formed, collapsed, and were beginning to fill with eolian and organic material.

Day 3

Enroute to Stop 13: Leaving Dinosaur State Park, we turn east on West St. In ¾ mile, turn right (south) on Rt. 99 (Main Street). In ¼ mile, cross Hampden Basalt ridge and then a small drumlin on right (See Stop 7 map, fig. 37). Drive down onto surface of Rocky Hill dam. In about ½ mile just south of the Rocky Hill-Cromwell town line, road dips down a bit and the entrance to the Dividend Brook spillway is to the right. Continue for 1½ miles over surface of unit **Mc** Lake Middletown delta deposits, part of the Rocky Hill dam. After traversing a couple of drumlins, descend off the delta and down onto the lake-bottom surface. In ¾ mile, get on Rt. 9 south, and go 1.2 miles to Stop light. Turn right for Rt. 17A north and Arrigoni Bridge. Cross the Connecticut River and stay straight on Rt. 17A north. Pass old Brownstone quarries to the left. Note many Portland brownstone buildings in the village of Portland. Continue on Rt. 17A for 3 miles to the junction with Rt. 17. Continue straight across onto Sage Hollow Road, than immediate left onto Cornwall Street. We will make a brief stop at this recently exposed section (see photo below) in unit **Mp** deltaic deposits of glacial Lake Middletown. The excavation exposes about 10 m of sandy subaqueous foreset beds that dip gently eastward. Coarser grained gravelly beds comprise the lower 2 m of the cut.



Returning to Rt. 17 north, we travel just above Connecticut River floodplain (to the west) for about ½ mile. After crossing Reservoir Brook, we ascend to the 49-m (160-ft) deltaic surface and travel along the contact with the till/bedrock hills to the east for about 2 miles. At Old Maids Lane, before turning left, note metamorphic bedrock outcrop on the right. Cross the eastern border fault of the Mesozoic Hartford basin (not exposed) and drive over the 50-53-m (165-175-ft) deltaic surface of the Rocky Hill dam. In about ½ mile we descend off the delta and drop down onto a stream-terrace surface. This 15-m (50-ft) surface records the breeching of the dam and lowering of the stable level of Lake Hitchcock by 3-4 m as the New Britain spillway was abandoned for this lower way out. In ¼ mile, we turn left on Tryon Street and in a tenth of a mile turn left into Glastonbury Bulky Waste Disposal facility.

Stop 13. Excavation at Town of Glastonbury bulky waste disposal site, 1145 Tryon Street, Glastonbury (-72.619, 41.636)

Figure 48. Gravelly, southwesterly dipping deltaic foreset beds overlying bottomset beds exposed in the Glastonbury Bulky Waste site excavation.

The excavation (fig. 48) is in an ice-marginal delta in the Cromwell deltaic deposits of glacial Lake Middletown (**Mc**, fig. 37). In this part of the river valley, the deposits are a series of ice-marginal deltas with surfaces at 50-56 m (165 to 185 ft). A topset-foreset contact at 45.5 m (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 deposits (**Mdb**, fig. 37), these deltaic sediments form a massive blockage at 47-53 m (155-175 ft) altitude in a narrow part of the Connecticut River valley and constitute the sediment dam for glacial Lake Hitchcock.

The relatively flat, noncollapsed delta plain is at 50 m (165 ft) in this area, but the bulky waste site excavation is cut into the 46-m (150-ft) surface of an ice-channel ridge at the juncture of walls of a compound kettle. There are 2 levels of excavation. Formerly, 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. Gravel clasts were boulder-sized, subangular to subrounded, with intermediate 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. 48) 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.

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 15.6 cal ka, the Rocky Hill dam was breeched as glacio-isostatic rebound began. Remnants of a stream terrace at 15 m (50 ft) in altitude cut through the dam deposits are preserved on both sides of the river in this vicinity (**st**, fig. 37). The 15-m terrace records the dam breeching event when the southern basin of glacial Lake Hitchcock drained and a river began to flow through this section of the Connecticut valley. This drainage only affected the part of Lake Hitchcock south of the Holyoke Range in Massachusetts.

Stop 14. Riverfront Park, Town of Glastonbury, 320 Welles Street, Glastonbury (-72.617, 41.712)

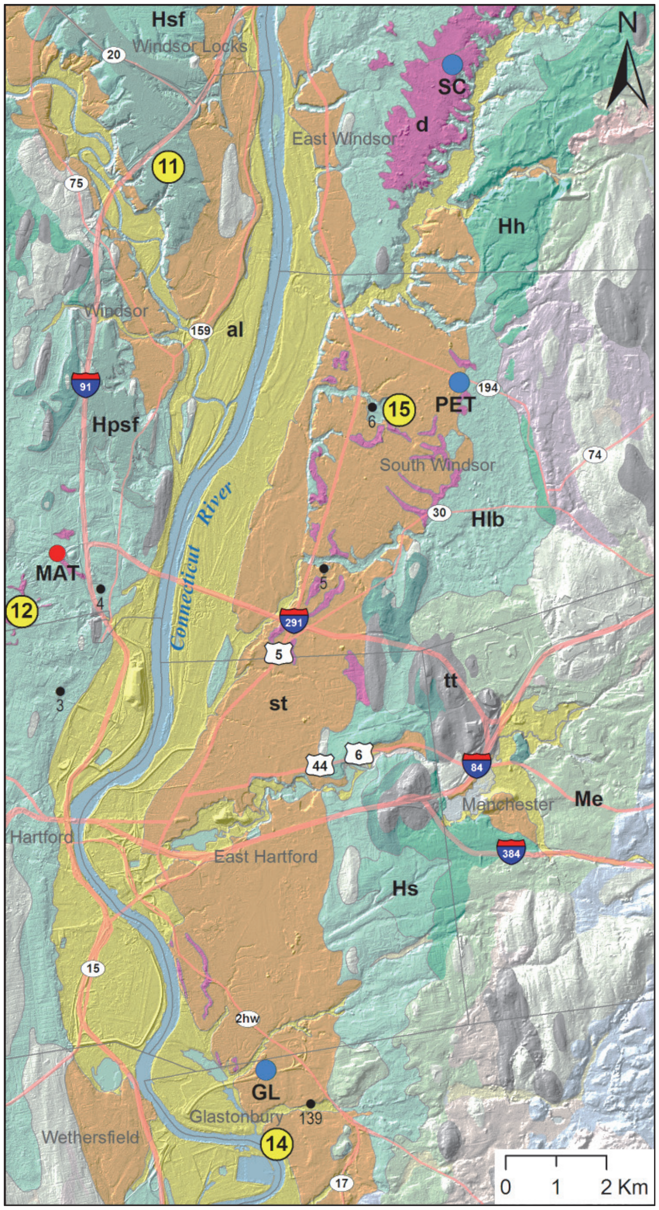


Figure 49. Quaternary geologic units in the vicinity of Stop 14 and 15 modified from Stone and others, 2005. **Me**—Lake Middletown deposits, **Hspf**—post-stable delta deposits, **Hh**—High-level deposits, **Hsf** and **Hs**—stable delta deposits, **st**—stream terrace deposits, **al**—floodplain alluvium. **GL**, **PET**, **SC**, **MAT**—Core locations.

In 2009, Jack Ridge and students collected 3 long cores in Connecticut to aid in the calibration of the New England Varve Chronology (see locations fig. 49 and varve plots fig. 50). We will use the Riverfront park location on the 15-m terrace surface along the Connecticut River, to lay-out and discuss sections of the varve cores. Currently, about 1.5 m of pebbly sand and gravel terrace deposits are exposed in a shallow excavation (fig. 49a) near the newly constructed Riverfront Boathouse. This location is about a kilometer south of the Glastonbury core site (GL, fig. 50) which is the southernmost of the 3 cores, and perhaps the most significant. Although the GL core did not reach the bottom of the varve section, it contains the single longest record measured in Connecticut—AM2868-4785, a total of 1,917 years. The core location is 10.5 km south of the Matianuck core site which yielded the oldest Connecticut basal varve—AM-2831. Based on the ice-retreat rate of 55 m/year indicated by other cores (fig. 2), we could expect there might be 190, or so, more varves below the Glastonbury core section.

The Glastonbury core sampled about 300 younger varves than had previously been known in Connecticut—providing evidence that varve sedimentation continued in Lake Hitchcock until at least 15,985 cal ka. The top of the varve section has likely been removed by stream terrace incision.

In August 2010, the Glastonbury core was featured in "Varves of the month" on the North American Glacial Varve Project website — as follows: [http://eos.tufts.edu/varves]

This month's varves are relatively thick red varves from a 152-ft core taken in the parking lot of the Putnam Bridge Center in Glastonbury, Connecticut near the Glastonbury/East Hartford town line. The varves are from a depth of 132 ft. and the split core has been partially dried to make the summer and winter layers more easily discernable. The varves on this image have been matched to the New England Varve Chronology and numbers on the image (2944-2952) are North American varve years. A plot of the outcrop core from which the image was taken and adjacent overlapping cores is also shown (fig. 51). The varves on the image were deposited at about ~17,900 yr ago in an environment 18 km south of the receding ice front when it was in East Windsor and Windsor Locks, Connecticut.

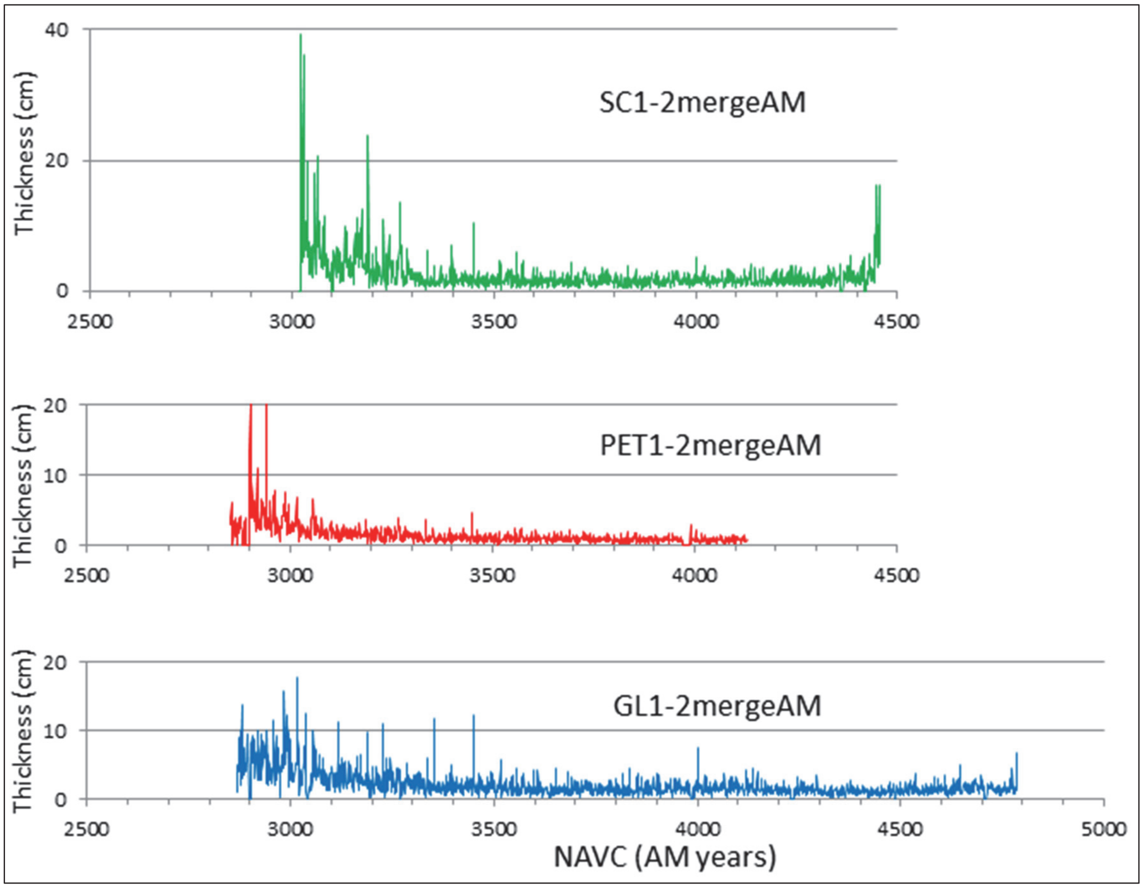


Figure 50. Plots of NAVC varves for three Connecticut cores

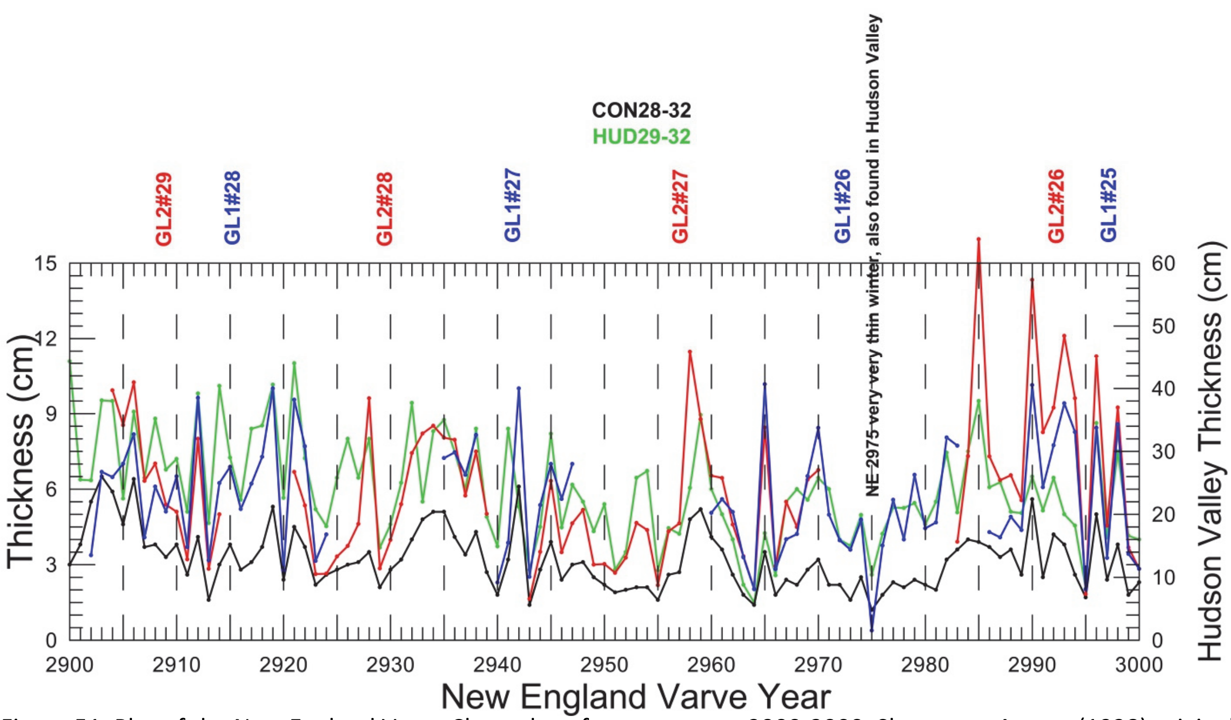
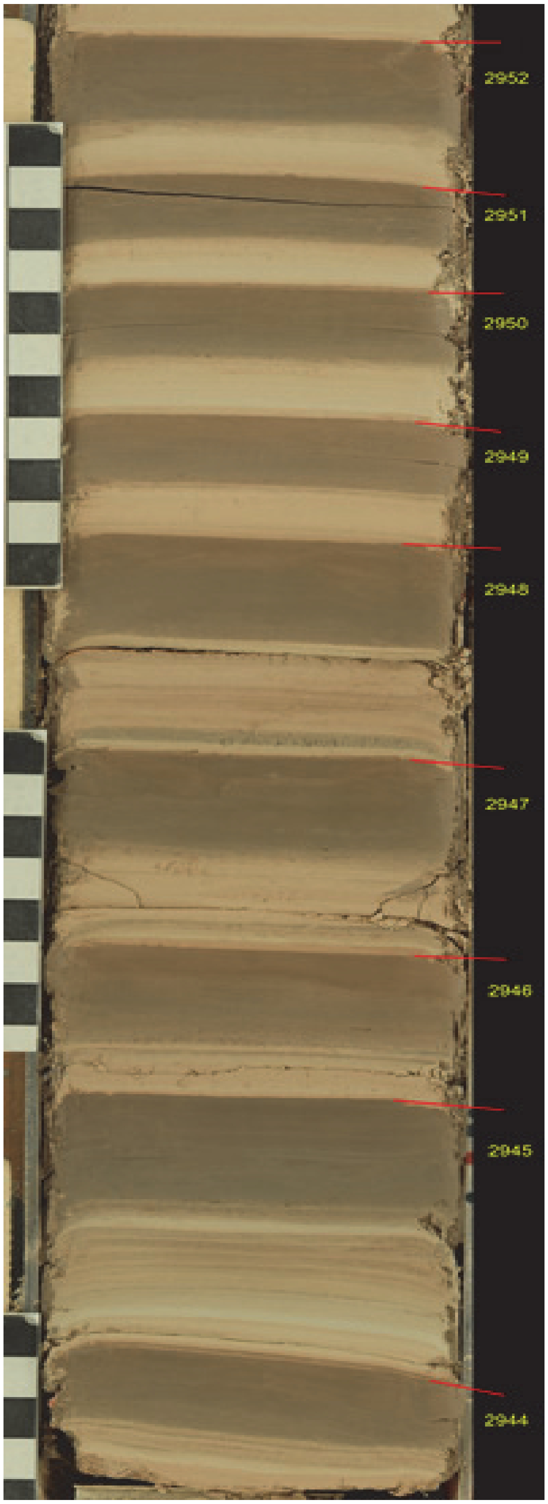


Figure 51. Plot of the New England Varve Chronology for varve years 2900-3000. Shown are Antevs (1928) original "normal curves" for Connecticut (black, CON29-32) and the Hudson Valley (HUD29-32) with matches of 5-ft core sections collected at the Glastonbury drill site including GL2#29-26 (red) and GL1#28-25 (blue). The image of varves below shows varves 2944-2952 in core GL2#27.



The varves shown here are typical of varves transitional from ice-proximal to distal with winter and summer layers having about the same thickness. There is a positive correlation between winter and summer layer thickness suggesting that: 1) winter layer clay (dark layers) was introduced to the lake during the summer and 2) time was not a factor in determining the thickness of winter layers. All of the summer layers have many laminations in them that represent melting or storm events during the summer and in some cases possibly diurnal cycles. Some of the varves, especially the thicker ones show thin early melt season layers but all the varves are dominated by their main melt season layers. The early melt season layer is sometimes expressed as a thin red or brown horizon at the bottom of each varve (see for example 2945, 2947, 2948, 2950 and 2951). In some cases this layer is nothing more than a red pigment in a silty layer, possibly the result of clay resuspended by the first melt season currents. Late melt season layers are not discernable except in varve 2945 (the thickest varve on the image) that also has a summer layer that becomes finer upward and shows a rhythmic, possibly diurnal, pattern in its upper part. There appears to be a relatively sharp change from summer to winter layers without the gradation often seen in thicker varves, except in varve 2947 where there is a gradational summer to winter transition. Winter layers in the varves show a very faint gradation in color from olive gray at the bottom to gray with a reddish brown tint near the top (see also last month's varves). This faint color change appears to be gradational and probably reflects grain size gradation (decreasing upward) in the winter layer with iron oxide pigments settling later in the winter than gray clay, which may have started to flocculate immediately after the melt season.

The varves also show a change from a relatively thick varve interval (2944-2948) to a thinner varve interval (2949-2952). On running average plots of varve thickness these thick and thin cycles are distinct and show sub-century scale changes in varve thickness in response to subtle changes in climate or weather patterns. Relatively warm periods are times of more meltwater production and deposition of thicker varves while cooler intervals were times of less meltwater production and thinner varve deposition. It is tempting to interpret varve thickness simply as a function of temperature, however, there are two factors that control the thickness of varves. Not only is mean annual temperature important but also the length of the melt season. Both of these factors contribute to the number of degree-days in the melt season.

STOP 15: KF Plant of Redland Brick Co., 1440 John Fitch Blvd., South Windsor, (-72.590, 41.844).
NOTE! The clay pits at the KF plant are an active mining operation and permission is required from the plant manager for entry.

Figure 52. Two of Jack Ridge's students, Jeremy Wei and Laura Carter, collecting a varve core at the KF Plant in 2009.

The Kelsey-Ferguson clay pits have been stops on sponsored field trips in the Connecticut Valley for several decades (Hartshorn and Colton, 1967; Ashley and others, 1982; Koteff and others, 1988; Stone and Ashley, 1992; Stone and others, 2005). By 2004 mining had exposed about 7 m of continuous varve section. The varves are light to medium olive gray with occasional pinkish to reddish gray tones and are 0.2-2.0 cm thick. The varves have well-defined winter clay beds and summer layers composed of multiple micrograded sandy to clayey silt beds. Summer layers also have the distinction of both starting and ending with coarse (fine to medium sand) layers that may represent overturning events at the beginning and end of the melting season. The varves also contain trace fossils, the most common of which is a small (~0.5 cm wavelength) sinusoidal trace produced by either insect larvae or nematodes (Ashley, 1972, 1975). A larger trace composed of evenly-spaced pits was discovered that may be the product of a fish (Benner and Ridge, 2004).

The gray varves represent a distal glacial varve sequence dominated by sediment derived from metamorphic rock sources that flank both sides of the Connecticut Valley. Near the base of the exposed sequence the varves begin to transition downward into red, hematitic varves with increased calcium carbonate concentrations as is evident from the many red concretions scattered across the floor of the clay pit. The lower red varves are dominantly derived from glacially eroded Mesozoic rocks of the Hartford Basin that underlie the Connecticut Valley. In the summer of 2004 an exposed face on the southeastern corner of the active pit was sampled with an overlapping set of 2-ft long, 3-inch diameter PVC cores for detailed study of the varve stratigraphy. A compiled varve sequence from the cores totaling 552 years was matched to Antevs' (1922) New England varve chronology (NE 3617-4168) without a single extra or missing varve in the section. Two radiocarbon AMS dates were obtained on organic material associated with varve NE-3826 and NE-4113 (Stone and Ridge, 2009). The dates are 14,120±80 (GX-32114) (16.9 cal ka) and 13,735±90 (GX-32113) (16.6 cal ka) respectively. In 2009, an additional 324 varves at the bottom of the section were sampled down to NA 3297 (NE 3293), bringing the varve total to 876 continuous varves.

In March 2011 KF Plant varves were featured as "Varves of the month"—description follows: Ridge, J.C. (May 11, 2015) *The North American Glacial Varve Project*. Retrieved from http://eos.tufts.edu/varves.

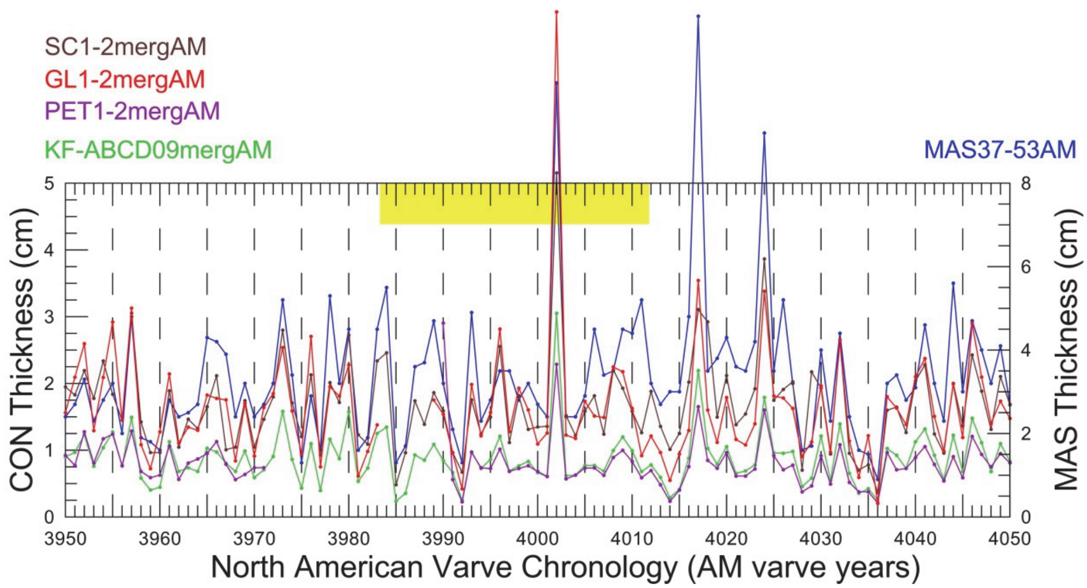
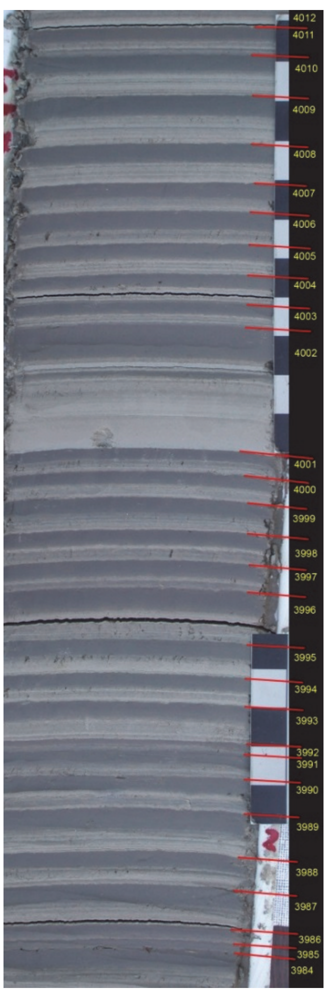


Figure 53. Plot of North American Varve Chronology varves AM 3400-3500 at sections in South Windsor (KF Plant - Redlands Brick Co. pit and Petersen Farm), and East Windsor (SC) and the Connecticut and southern Massachusetts records of Antevs (1922). The yellow swath is the part of the plot shown in the core image below. Note that the scale for Massachusetts varve thickness (right side) is different and the Massachusetts varves are much thicker than the Connecticut varves as a result of being closer to the receding ice front when they were deposited.



Varve thickness was largely controlled by glacial meltwater activity as meltwater moved south away from the glacier but at least some of the sediment in the varves was entrained by glacial meltwater from local tributaries that were non-glacial at the time of varve deposition. Varves in this part of the sequence are primarily gray (unlike varves lower in the section that are red) and have only a very faint red tint in the clay beds of some varves (AM 3467-3469 and 3489-3490). This indicates that meltwater entering the lake from the glacier and meteoric sources were no longer contributing significant red sediment to this area of the lake. In the slightly red-colored varves the red color is in the winter clay bed and primarily near the top of the clay units. This may mean that the red component of the sediment (probably a hematite pigment) is finer than most of the clay mineral particles making up the lower and grayer part of the winter layer.

The summer or melt season layers of each varve (light-colored units) are composed of a stack of micrograded units of fine sand and silt (especially for example AM 3469-3472, 3477, and 3491-3496), although there are not as many distinguishable units as lower in the section where the varves are much thicker and strictly ice-proximal. The upper parts of some summer layers may have a slightly grayer appearance than lower in the summer layer and many summer layers grade into the winter layer above. Summer layer thickness in this sequence shows a strong positive correlation to winter layer thickness. This suggests that winter layer thickness is controlled by the volume of clay introduced to the lake during the summer and is dependent, like summer layer thickness, on summer sediment discharges to the lake.

Figure 54. Stream terrace deposits capping varves at the KF Plant.

About 5 m of fluvial, cross-bedded, pebbly sand unconformably overlies the silt and clay varves of glacial Lake Hitchcock in the clay pit area (fig. 55). Water draining from Lake Hitchcock north of the Holyoke range in Massachusetts flowed across exposed lake floor in the southern basin. The cross-bedded sands capping the varves at the KF Plant represent first, erosion and then, deposition by a fluvial drainage system that began with the initial failure of the Rocky Hill dam and continued as glacio-isostatic uplift progressed. The terrace provided a source of sand for the well-developed arcuate dunes traversed by Foster Road southeast of the clay pit. Much more subtle dune ridges can be seen in the high-resolution lidar image (fig. 55).

To the east of the dunes, the lake-bottom surface is dotted with interfering forms of rimmed depressions that are interpreted as lithalsa scars.

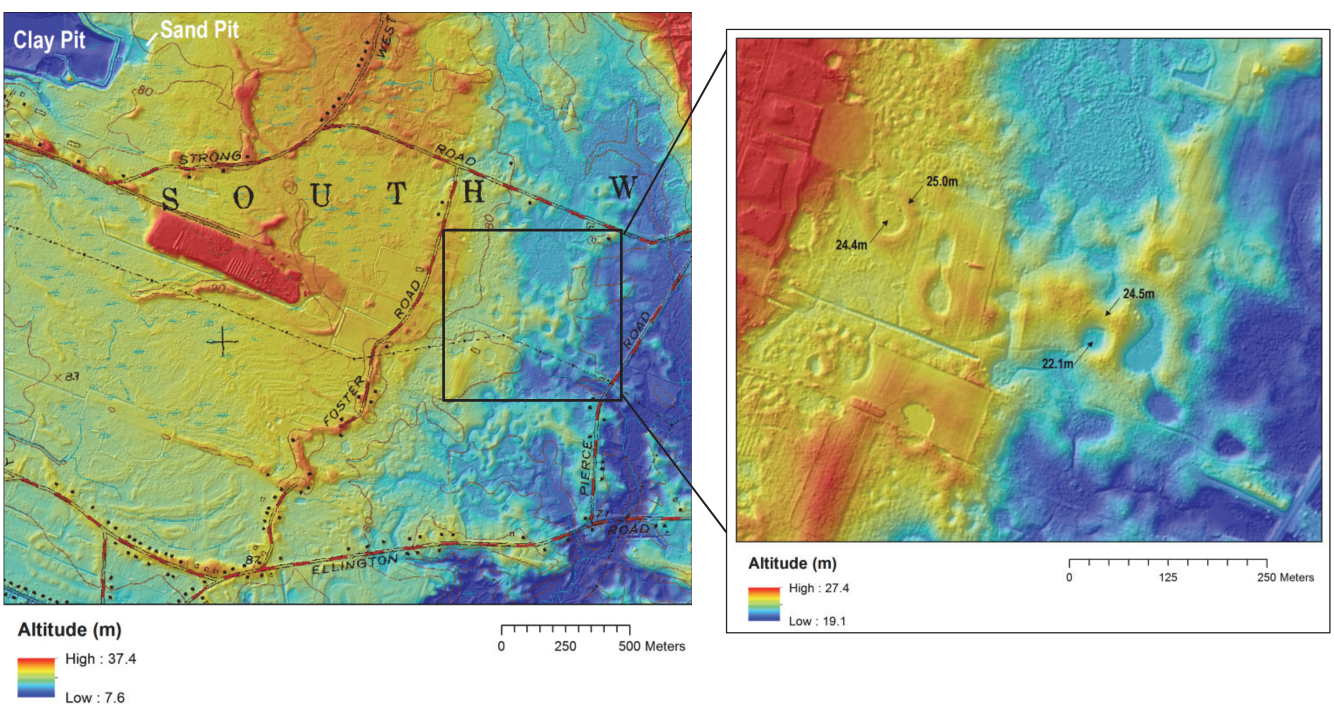


Figure 55. Lidar-generated DEM (vertical resolution 9.25 cm) in the area southeast of KF Plant clay pit.

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- ★ Jack Ridge
- ★ Ralph Lewis
- ★ Mary DiGiacomo-Cohen
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