Quaternary International 342 (2014) 173–185

Contents lists available at ScienceDirect

Quaternary International

journal homepage: www.elsevier.com/locate/quaint

Hydraulic back-flood model for the archaeological stratigraphy of the Connecticut River Alluvial Lowland, central Connecticut, USA

Robert M. Thorson [a,b,*], Daniel Forrest [c], Brian Jones [d]

[a] Center for Integrated Geosciences, University of Connecticut, Storrs, CT 06269, USA
[b] Dept. of Ecology & Evolutionary Biology and Dept. of Anthropology, University of Connecticut, Storrs, CT 06269, USA
[c] State Historic Preservation Office, One Constitution Plaza, Hartford, CT 06103, USA
[d] Archaeological & Historical Services, Inc., 569 Middle Turnpike, Storrs, CT 06268, USA

ARTICLE INFO

Article history:
Available online 26 April 2014

Keywords:
Floodplain
Stratigraphy
Archaeology
Connecticut U.S.A.
Hydrology

ABSTRACT

State-mandated archaeological investigations associated with urban renewal in downtown Hartford led to the development of a mechanistic model for Holocene floodplain sedimentation in central Connecticut. Our model is based on: historic flood hydrometeorology and geomorphologic mapping; the lithologic, magnetic, pollen, and archaeological stratigraphy exposed in sixteen deep boreholes and 24 hydraulic push cores; the chronology provided by sixteen AMS radiocarbon ages, and diagnostic historic artifacts ranging from 10,030 cal BP to the present. We conclude that the alluvial stratigraphy in this part of the lowland resulted from bottom-up changes in hydraulic ponding at a bedrock outlet, rather than from top-down responses of the watershed to changes in climate, vegetation or human activity. Our model provides a geologically based time-space framework for the distribution of known archaeological sites, and carries implications for future research.

1. Introduction

Archaeological excavations, especially those associated with cultural resource management projects (CRM), usually begin with the discovery of artifacts *in situ*, then proceed by fitting these cultural materials into a regional framework. The Adriaen's Landing Project (ALP) gave us an opportunity to work in the opposite direction. There, a massive construction project was being planned for an urban "brownfield," an asphalt-paved historic floodplain contaminated with toxic waste. Because a traditional CRM reconnaissance survey was impossible, funding was wisely steered toward the development of a stratigraphic model that could be applied to the entire Connecticut River Alluvial Lowland (CRAL), here defined as the channel, floodplain, and postglacial fluvial terraces within an alluvial reach between two bedrock pinch points (Fig. 1). Our initial work piggybacked on the first round of geotechnical investigations, which involved standard split-spoon sampling from hollow-stem augers. After encouraging initial results, we returned to collect a series of strategically placed GeoProbe (hydraulic push) cores that allowed continuous sampling for lithostratigraphy, pollen, and radiocarbon dating to a depth of 11.0 m (36 ft).

The ALP can also be viewed as a case study of mistaken assumptions. Based on both its location within the Holocene floodplain and on previous research on alluvial deposits both above and below the CRAL (Jahns, 1947; Patton and Horne, 1991; Stone et al., 2005), and we expected our cores to encounter facies characteristic of meandering rivers, for example, channel, bar, levee, overbank, and backswamp (Brakenridge, 1988; Gladfelter, 2001; O'Connor and Webb, 1988). We also hypothesized that the vertical stacking of alluvial deposits would reflect changes in upriver catchment processes (Schuldenrein, 2003). Instead, we learned that the graceful curve of the river below our project area was not a meander bend, and that the overbank strata we expected to find turned out to be, in effect, lacustrine sediment deposited in transient flood lakes (up to 4 km wide, 30 km long, and lasting up to nine days in duration).

The purpose of our project was to develop a stratigraphically based archaeological model for the CRAL. The purpose of this paper is to highlight for the geo-archaeological community our hydraulic back-flood mechanism, which will likely apply to other comparable settings. Accordingly, we: (1) describe the regional setting; (2) subdivide the CRAL into distinct reaches; (3) document the

* Center for Integrated Geosciences (U-1045), 354 Mansfield Road, Storrs, CT 06269-1045, USA
E-mail address: robert.thorson@uconn.edu (R.M. Thorson).

http://dx.doi.org/10.1016/j.quaint.2014.03.026

backflood mechanism with evidence from historic floods; (4) present the borehole stratigraphy; and (5) draw archaeologic implications.

2. Setting

2.1. Geologic

The Connecticut River flows southward from hard-rock highlands at the New Hampshire–Quebec border to tidewater at Long Island Sound, making it the longest river in New England. Its northern half, flowing roughly parallel to the Vermont–New Hampshire border, follows a suture zone of early Paleozoic age. Within Massachusetts and Connecticut, it flows through lozenge-shaped Mesozoic rift basins that opened as half-grabens during the initial breakup of Pangea in the early Mesozoic. Below Middletown, CT the river turns to the southeast, flowing discordantly across the rock structure, but parallel to the regional slope. There, the Connecticut River is deeply entrenched through Proterozoic crystalline rocks, flowing through what is locally referred to as its "canyon" that is tens of meters deep. In the river's final reach below Essex, it flares broadly into a tidal estuary that indents the north shore of Long Island Sound at Old Saybrook.

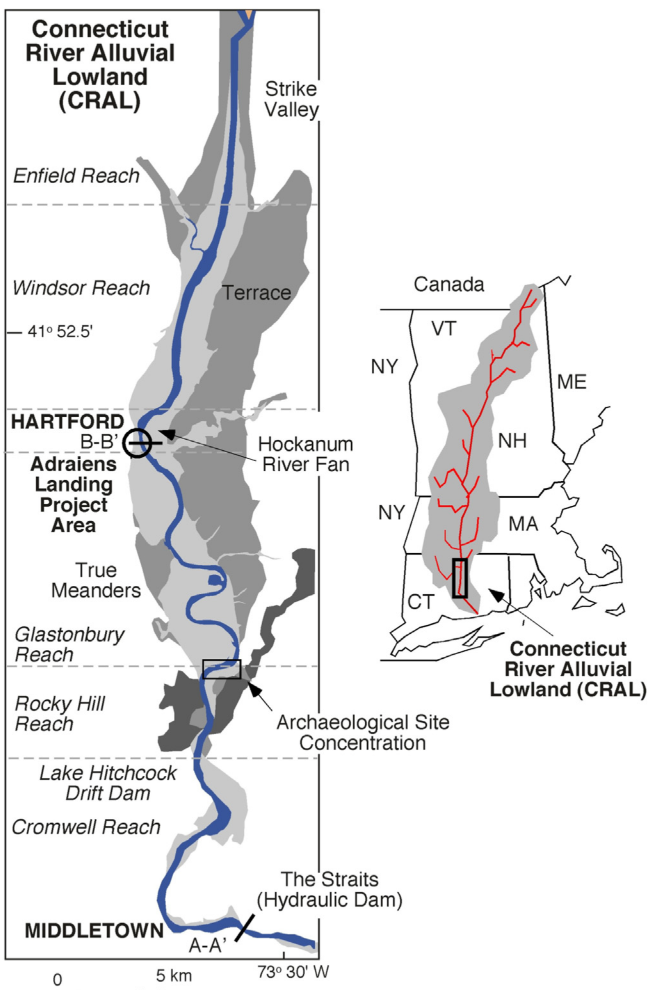

Fig. 1. Reaches of the Connecticut River Alluvial Lowland (left) and its location within the Connecticut River watershed (right). Circle indicates location of the Adriaen's Landing Project Area. Lines A–A' and B–B' show locations of valley cross sections in Fig. 2. Box in Rocky Hill Reach is a concentration of sites described in the text. Selected features are named on the map. The drift dam of Glacial Lake Hitchcock (dark gray), a high-level, post-lake alluvial terace (medium gray), and the modern floodplain (light gray) are mapped.

The glacial history of the study area was governed largely by its pre-existing bedrock structure. The mass flux was higher within the lowland associated with the Mesozoic rift than on the adjacent highlands, concentrating subglacial meltwater, enhancing erosion, and deepening the basin to well below sea level. Deglaciation took place about 22 ka cal yr BP (Ridge, 2003), initially under very cold conditions (Peteet et al., 2012). At this time, global sea level was approximately 120 m lower than present, and the thick rigid crust of New England was glacioisostatically depressed, or flexed down to the north as a result of ice-loading (Peltier, 1996). As a result, the Connecticut River "canyon" was plugged with a combination of coarse glacial debris and sandy meltwater deltas, one of which formed the Rocky Hill sediment dam, which diverted river discharge through stable bedrock spillways. The net effect was to impound Glacial Lake Hitchcock, the longest glacial lake known to have existed in New England, and which covered the entire area now occupied by the CRAL. It began to form as the ice front retreated to the current Massachusetts/Connecticut state line between 17 ka and 16 ka, reaching a long stable stage around 15 ka, followed by drainage of the lake basin south of the Holyoke Range by 13.5 ka, and drainage of the northern sections by 12.5 ka (Thorson and Schile, 1995; Rittenour et al., 2000; Ridge, 2003; Stone et al., 2005).

A thick veneer of varved silts and clays was deposited on the floor of Glacial Lake Hitchcock. A series of massive sandy deltas were graded to its shorelines. After the lake drained, and driven by a combination of isostatic rebound to the north and the erosion of the sediment plug from the canyon, the former floor of Lake Hitchcock was eroded. This produced a flight of broad fluvial terraces up to 7 km wide capped by sandy outwash deposits. The sequence requires: an early phase of slow downcutting responsible for creating the broad tread; deep entrenchment caused by a combination of lake drainage events further upstream, isostatic rebound, and a reduction in sediment supply within a reforesting watershed; and aggradation of the Holocene valley fill, which had begun in Massachusetts by ∼9 ka (Stone et al., 2005). Holocene changes in upland settings were more closely associated with climatic changes (Bierman et al., 1997).

2.2. Flood hydrometeorology

Upstream from the CRAL the watershed increases in local relief along a pair of north–south divides. To the west are the Green Mountains in Vermont, and their more subdued equivalents to the south, the Berkshires and Litchfield Hills in Connecticut. To the east are the White Mountains of New Hampshire and a ridge-forming anticlinorium in Massachusetts and Connecticut to the south. These parallel highlands rise northward on both sides of the Connecticut River, forming the most rugged terrain in New England, with hundreds of peaks more than 1 km high.

The catchment extends from 41 to 45° latitude as it rises in altitude from the broad-leaved deciduous forests of the Atlantic coastal plain to boreal mixed conifer hardwoods of the interior highlands. Seasonal snowfall totals range from nearly zero on the shore of Long Island Sound to several meters to the north. Lake and river ice are annual occurrences, thicker before the onset of warming in the late 20th century (Solomon et al., 2007). The north–south orientation of the catchment parallels the onshore direction of the distal tracks of hurricanes and tropical storms, which rain themselves out in that direction.

The orientation, length and ruggedness of the catchment, combined with its nearly complete forest cover, thick snowfall, and proximity to North Atlantic moisture give rise to a runoff hydrology in the CRAL dominated by flood stages that rise and fall slowly in response to: (1) late-summer and early autumn storms originating in the subtropics (hurricanes), or (2) snowmelt associated with heavy rain in spring (Patton, 1988). This distinctly bimodal seasonal runoff hydrology is more pronounced than that of the Hudson River. Catchments to the west (i.e., the Delaware, and Susquehanna) are less dominated by snowmelt hydrology and less likely to receive hurricane precipitation. Catchments to the east (i.e., the Quinebaug, Merrimack, Saco, and Kennebec) are shorter and oblique to the tracks of subtropical storms.

The postglacial paleoclimate is fairly well documented for New England (COHMAP, 1988; Shaw and van de Plassche, 1991; Thorson and Webb, 1991; Webb et al., 1993; Thorson and Schile, 1995; Briffa et al., 2001; Peteet et al., 2012; Pedereson et al., 2005). Regional records indicate: (1) cold and moist tundra prior to about 12.5 ka; (2) an abrupt short-lived postglacial transition to dry cold conditions; (3) a return to moister conditions during the Younger Dryas; (4) several millennia of drier conditions during the so-called Anathermal; (5) early Holocene insolation maximum; (6) a broader mid-Holocene interval of warmth, general dryness, and climatic stability, the so-called "Hypsithermal" between about 8 ka and 3 ka; and (7) a return to cooler and wetter conditions during the last three millennia with a post-contact period of Euro-American settlement cooler and wetter than the majority of the present century. Though these regional climate variations are critical to river discharge models based on water balance (Knox, 1993), they exert only a weak influence on the peak floods responsible for the bulk of alluvial sediments constituting the ALP stratigraphy, which are dominated by extreme runoff events, and stochastic factors such as transient dams due to river ice or log jams.

2.3. Holocene sea level

Breaching of the Rocky Hill sediment dam for southern Glacial Lake Hitchcock took place as early as 13.5 ka (Lewis and Stone, 1991; Stone et al., 2005). At that time, sea level was approximately −40 m below modern, based on correlation with a submerged delta at the mouth of the Connecticut River in Long Island Sound (Gayes and Bokuniewicz, 1991). Also, the thalweg of the Connecticut River was isostatically depressed by an unknown amount at this time. By the time floodplain aggradation began at the Adriaen's Landing Project study area (ALP), most of the uplift was complete and the rate of sea level rise was decelerating. Stone et al. (2005) reconstruct the 9 ka sea level at −30 m, and the 6 ka sea level at −5 m, yielding an average submergence of 8 mm/yr. Deceleration continued into the late Holocene. Based on *in situ* stumps and marsh peat dating between 4.5, and 1 ka, Patton and Horne (1991) reconstructed a much slower submergence rate of about 1.7 mm/yr for the Connecticut River estuary. The presence of salt marsh deposits at the mouth of the Connecticut River near Guilford indicates that the transgression continued to decelerate until about 1200–1300 A.D., with slightly higher rates since then (Varekamp et al., 1992; Van de Plassche, 2000).

3. Connecticut River Alluvial Lowland (CRAL)

3.1. Reaches

The CRAL is delineated on the east and west by lacustrine deposits of Glacial Lake Hitchcock (and its predecessor, Glacial Lake Middletown) and the deltas and moraines above their shores. On the north and south it is delineated by bedrock constrictions. The upstream constriction occurs where the river enters a narrow strike valley within the Jurassic strata. The downstream constriction, known as the Straits, is the most confined cross-section within the canyon (Figs. 1 and 2). It occurs where the Connecticut River crosses the Eastern Border Fault, which juxtaposes strong Proterozoic rocks against weak Mesozoic redbeds of the rift basin.

For archaeological purposes, the CRAL is divided into five specific reaches, each delineated by a change in fluvial processes. In the *Cromwell Reach*, just above the Straits, the Connecticut River is deeply entrenched into Mesozoic sedimentary rocks. Next is the *Rocky Hill Reach*. There it flows through a massive plug of coarse-gravel delta moraines known as the Rocky Hill Dam, the major blockage for Glacial Lake Hitchcock. This dam coincides with an abrupt narrowing of the valley located where east-tilting basalt ridges are truncated by the Middletown Fault. Floodplain segments within these reaches are localized and narrow.

The *Hartford Reach* is fully urbanized and guarded by high artificial levees. The westernmost point of the Connecticut River in this reach is the deepest (8.5 m; 28 ft) and narrowest (155 m; 508 ft), which is why this site was chosen for the earliest bridge. Engineering boreholes indicate that the river is cutting against a submerged prong of sandstone bedrock bounded by a normal fault that descends from just below the ground surface west of the ALP, to a depth of more than (30.5 m; 100 ft) on the east side of the River (Handman and Hildreth, 1972). Opposite the bedrock cliff is a bend in the river parallel to the western edge of a fan-shaped terrace remnant (Stone et al., 2005). This remnant lies opposite the largest tributary in the vicinity of the Hockanum River. We interpret this suite of features to indicate that the channel of the Connecticut River marks the outer limit of a broad alluvial fan composed of reworked glacio-deltaic sediments, and built by an important tributary (the Hockanum River) in response to entrenchment by the Connecticut River. This fan was later veneered by Holocene alluvium exhibiting no morphologic or stratigraphic evidence for channel migration by the Connecticut River. Instead, a permanent overflow channel across the buried Hockanum River alluvial fan that resembles a meander chute cutoff (now blocked by artificial dikes).

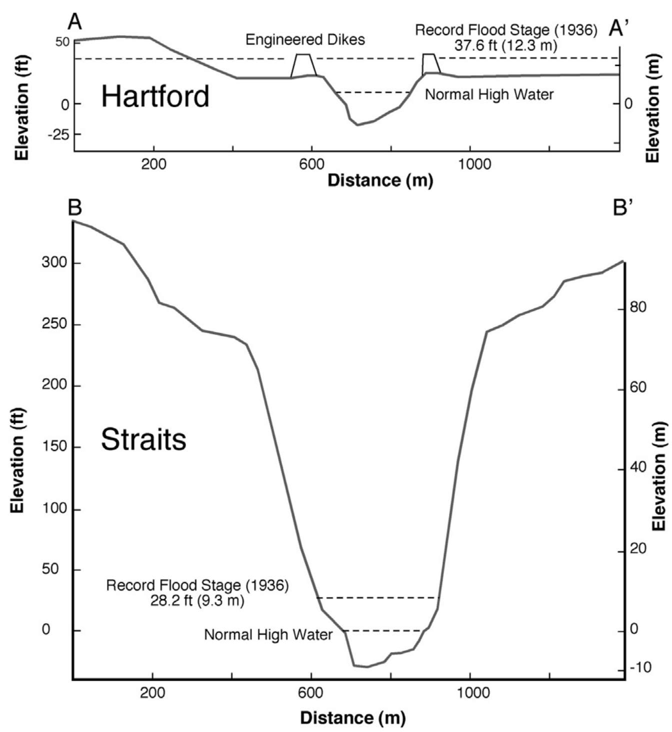

Fig. 2. Valley cross-sections for the Adriaen's Landing Project area (A–A') and the Straits (B–B'). Elevations are relative to the North American Vertical Datum.

The southern edge of the Hartford Reach is defined by a flow expansion just upstream from the mouth of the Park River. This confluence with the main stem is likely fixed in place above a bedrock paleo-valley (Handman and Hildreth, 1972). Into this flow expansion, and prior to the 20th century when it was channeled underground, the Park River brought copious sand and gravel into the main channel, forming the historic Hartford bar, a hydraulic steady-state phenomenon. This was the head of navigation of most craft prior to 19th century dredging and channel improvements. The heavy and coarse-grained sediment flux associated with the Park River is historically documented (Houston, 1868). For example, when it was dredged in 1884, 83% of the sediment returned within a year.

The Hartford Reach divides the central CRAL into the *Glastonbury Reach* to the south and the *Windsor Reach* to the north. Only in the Glastonbury Reach does the river freely meander on its floodplain. Only two-and-a-half free meanders are present with an average wavelength of 4.1 km, an amplitude of 1.9 km, and a radius of curvature of 0.9 km, respectively. Rapid and active channel migration is indicated by fresh meander scroll topography, historic mapping, and eyewitness accounts of bank loss. Based on re-mapping by the U.S. Geological Survey between the 1890s and 1969–1980, bank migration has exceeded 200 m. Within the Windsor Reach to the north, the channel is fairly straight and stable, with a pattern dominated by parallel levees, Yazoo streams, and backswamps (Metzler and Damman, 1985). This reach has two large, quasi-stationary gravel islands (lateral bars, now fully forested) on opposite sides of the river at the mouths of the Farmington and Podunk Rivers. There are no bedrock outcrops within the reach.

Farthest north is the *Enfield Reach*, a narrow, strike valley about 7 km long following a non-resistant bed of the Portland Formation. A series of rapids and low waterfalls on bedrock were present in the 19th century, though they were modified by changes associated with mill construction.

4. Flood mechanism

Historic flood records for the CRAL summarized by Kinnison et al. (1938) for the interval 1639–1936 and extended forward in time by Thomson et al. (1964). Measured flows for the 162-year interval between 1843 and 2004, supplemented by stage records for unusually high floods dating back to 1683, were summarized by Ahearn (2005). In both analyses, the flood of 1936 was a strong outlier on the historic flood series, reaching a record stage of 37.6 ft (11.46 m). Major floods after 1936 likely constitute a series with a shorter recurrence interval, perhaps owing to a reduction in channel volume caused by sedimentation, and to massive engineering projects including bank protection, diking, channelization, and the construction of sixteen enormous flood-retention reservoirs. Because our study is archaeological in nature, we ignore the flood records after 1936 as not being representative of the Holocene pattern.

Between 1801 and 1936, there were 155 floods at or above the flood stage of 11 ft (3.3 m), based on the North American Vertical Datum of 1927. The paved surface at Adriaen's Landing in 1999 averaged about 10 ft (3.1 m) higher on the same datum (21.38 ft; 6.5 m), based on careful survey at four boreholes, an interval roughly equivalent to the thickness of the artificial fill. Forty-three of these historic floods were high enough to have inundated this modern surface, yielding an average recurrence interval of 3.14 years. All but eight of the 155 floods would have inundated the pre-fill surface, meaning that submergence was nearly an annual event. Though inundation took place during every month of the year, most common were spring floods (the historic vernacular term was "freshet") dating from late March through late April, when warmth-driven snowmelt coincided with heavy rain. Other contributions to significant flooding include summer storms associated with hurricanes and tropical storms migrating up the coast and November storms associated with extratropical cyclones known as Nor'-easters, also caused significant flooding. The decades between 1870 and 1890 stand out for their lower incidence of flooding.

The normal flood mechanism for large alluvial rivers occurs when the rising crest of a flood wave moving downstream locally overtops its channel (Wolman and Leopold, 1957). Water usually spills out first through a saddle or breach in a natural levee. From there it spreads out as a sheet of turbulent, sediment-laden water flowing over the floodplain. Inevitably, such breaching causes channeling in channel-proximal settings, surface scour in medial zones, and slackwater deposition in distal areas; multiple flood surges within single events lead to greater complexity. Surface scour also occurs during flood recession, though it is usually less significant. The pattern for higher parts of the floodplain does not involve erosion, but slow inundation followed by slackwater settling of fines. This mechanism operated during historic time for the floodplain north and south of the CRAL (Wolman and Eiler, 1958). It was especially well documented in Massachusetts for the 1936 flood by Jahns (1947). Based on 611 field measurements on the Connecticut River floodplain in western Massachusetts, an average of 3.5 cm of fresh sediment dominated by fine-to-medium sand was spread across the entire floodplain. Up to 25.9 cm of coarser sand was deposited on levees, with 13.2 cm on flat surfaces near the river. Local erosion was common and intense, creating conspicuous surface channels and scour pits below stationary eddies that were later backfilled with mud.

This normal flood mechanism, however, does not operate within the CRAL south of the Windsor Reach. There, flooding does not occur as an outward and forward spill from the crest of the flood-wave, but as a slow, up-valley submergence caused by backflooding above a hydraulic dam at the Straits. The flaring cross section of the bedrock valley at the Straits approximates that of a typical V-notch weir. Using the Cone equation, the stage height (h) in feet is controlled by the incoming discharge (Q) in cubic feet per second:

$$h = (Q/2.49)^{-2.48}$$

With a negative root greater than unity, higher and higher flood discharges are accommodated by progressively smaller changes in stage, giving rise to what is, in effect, an asymptotic upper limit to the height of a transient lake. This asymptote was confirmed for historic floods at the Straits by precise leveling of "natural" flood flows (Houston, 1868). For the six strongest flows, the difference in peak stage was less than 1.5 m. For a river with a normal bed slope, the effects of hydraulic damming would be limited to the first few kilometers upstream. The downstream gradient of the CRAL, however, is negligible (5–12 cm/km; $<1 \times 10^{-4}$) because it is controlled not by the channel thalweg, but by the viscosity of the water for any given temperature and the phase of the diurnal tidal cycle, whose mean rise and fall at Middletown and Hartford is 1.93 and 1.7 ft, respectively. So low is the gradient of the lower Connecticut River that tides are felt at Windsor Locks, nearly 20 km north of the ALP and 92 km upriver from Long Island Sound.

The combination of a resistant bedrock notch at the Straits, negligible river gradient, and a typical week-long flood duration converts the CRAL into a back-flooded lake, the volume and duration of which are controlled by the balance between inputs through the Enfield Reach and outputs through the hydraulic dam. Additionally, the dominant source for suspended sediment in floodwaters is the lacustrine fill of Glacial Lake Hitchcock, the silt and clay constituting the bulk of the river's banks for more than 400 km upstream. In other words, floodplain sedimentation at the ALP is a lacustrine process dominated by re-deposition of lacustrine sediment.

The effects of hydraulic ponding were best recorded for the flood of record, the vernal flood of March, 1936 (Fig. 3), which had a peak discharge of about 7560 m^3 sec^{-1} (270,000 cubic feet per second). Three lines of evidence indicate hydraulic ponding: (1) the extremely low measured surface-water energy slope, even at times of peak flow; (2) the counter-clockwise time loop (hysteresis) shown by a plot of stage vs. discharge requires a backup of water, first within the channel between March 15 and March 18, and then within a much larger temporary reservoir between March 18 and March 27; and (3) peak stage at Hartford occurred long after feeder tributaries were subsiding, indicating that constriction at the outlet was the controlling mechanism. The previous flood of record in 1854, though seven feet lower in stage, inundated the floodplain with up to 4.9 m (16 ft) of water for nine continuous days. This duration also indicates significant hydraulic ponding.

The rise of Holocene sea level at the Straits, and the associated infilling of the thalweg with resistant coarse gravel, would have progressively blocked the bottom of the hydraulic dam. Each vertical increment of rise at the bottom, whether by sediment or standing water, removed cross-sectional area that would have formerly been available to transmit flood discharge. Also, every increment of rise would have been associated with a proportionately smaller rise in peak stage, owing to the greater cross-sectional area at the top of the V-notch canyon per unit height, and to diminished bank friction. Hence, for any constant rate of sea level rise, the onset of hydraulic ponding will more dramatic than the subsequent rate of rise. Additionally, the volume of the transient reservoir holding the lake must diminish as it is filled from the bottom up with sediment, meaning a rise in the height of the "floodplain" constitutes a loss in lake volume. These relationships require that, for any given sedimentation rate, the average down-valley velocity of floodwave through the transient lake must decrease early in the history of the reservoir when peak stage is rising faster than floodplain aggradation, then increase when sedimentation raises the elevation of the floodplain faster than the asymptotic rise of peak stage.

Before the onset of hydraulic ponding, the Connecticut River in the CRAL would have been strictly alluvial in behavior, its hydraulic geometry and bed-bank materials adjusted to the dominant (i.e. channel-forming) flood discharge, and with a recurrence interval on the order of several years. The next stage would have been transitional, during which scour and deposition caused by overbank flows would have been arrested by backflooding associated with delayed rise of the transient reservoir. The final stage, after hydraulic ponding was well established, would have been strictly lacustrine. In this stage, deposition of clay and silt from slackwater suspension would have raised the "floodplain" above the level set by channel hydraulic geometry, creating an "underfit" river no longer able to overtop its banks.

5. Stratigraphy

5.1. Methods

The ALP began in 1999 with stratigraphic monitoring of sixteen geotechnical boreholes equipped with split spoon samplers, each 24 inches long and 2 inches diameter (Figs. 4 and 5). Samples from five boreholes were examined for lithostratigraphic, pollen and phytolith analysis, and sub-sampled for radiocarbon dating (Table 1). The main results of this phase were the characterization and dating of stratigraphic units at the deepest levels. Units 1 (till) and 2 (lacustrine clays) were restricted to these cores.

Table 1
Radiocarbon dates from Adriaen's Landing Cores.

Elevation[a]	Core		Sample				Analysis		Conversion		
m asl	Type**	Number	Unit#	Unit name[d]	Material	Beta lab#	Measured date	$^{13}C/^{14}C$ Ratio	Conventional age	Calibrated date[b]	Calibrated date BP
3.95	G	R5g	8	Organic Silt	Charcoal	Beta-154110	1060 ± 40	−25.3	1060 ± 40 BP	AD 990	960
3.66	G	R3g	8	Organic Silt	Charcoal	Beta-154105	1010 ± 40	−26	990 ± 40 BP	AD 1020	930
3.25	G	F15	8	Organic Silt	Charcoal	Beta-154102	1130 ± 50	−24	1150 ± 50 BP	AD 890	1060
2.75	G	P2	7	Massive Silt (U)	Wood (twig)	Beta-154099	1700 ± 50	−25.2	1700 ± 50 BP	AD 370	1580
2.52	G	F6	7	Massive Silt (M)	Wood	Beta-154101	1630 ± 40	−26.8	1600 ± 40 BP	AD 430	1520
1.75	G	P2	7	Massive Silt (M)	Charcoal	Beta-154095	2680 ± 40	−25.5	2670 ± 40 BP	820 BC	2770
1.74	G	S3	7	Massive Silt (L)	Charcoal	Beta-154104	3060 ± 40	−25.9	3050 ± 40 BP	1310 BC	3260
−2.29	S	16D	6	Mottled Mud	Wood	Beta-136488	6380 ± 50	−28.7	6320 ± 50 BP	5305 BC	7255
−2.29	S	16D	6	Mottled Mud	Wood	Beta-136487	6440 ± 40	−28	6390 ± 40 BP	5350 BC	7300
−2.27	G	P2	4	Muddy Sand	Wood-flotsam	Beta-154094	7870 ± 40	−24.6	7880 ± 40 BP	6680 BC	8630
−2.43	G	P2	4	Muddy Sand	Wood	Beta-154108	7960 ± 40	−24.6	7970 ± 40 BP	6920 BC[c]	8870
−3.08	G	R5	4	Muddy Sand	Wood-flotsam	Beta-154103	7870 ± 60	−28.2	7820 ± 60 BP	6650 BC	8600
−3.10	G	P2	4	Muddy Sand	Leaf litter	Beta-154107	7900 ± 50	−23.3	7930 ± 50 BP	6790 BC	8740
−3.20	S	(HA-60)	4	Muddy Sand	Wood	Beta-136490	8030 ± 40	−27.8	7990 ± 40 BP	7035 BC	8985
−3.22	S	(GZA 14-H)	4	Muddy Sand	Charcoal	Beta-137721	8130 ± 40	−24.4	8140 ± 40 BP	7080 BC	9030
−3.26	S	(GZA 15-G)	4	Muddy Sand	Wood	Beta-136486	8920 ± 40	−25.9	8900 ± 40 BP	8080 BC[c]	10,030

** G stands for Geoprobe sample, S stands for split spoon sample.
[a] Samples arranged by stratigraphic unit from top down, then elevation. Note concordance.
[b] InterCAL 98 database.
[c] Average of multiple intercepts.
[d] Refer to text for details.

In the next phase during the 2001 field season we chose our own coring sites, taking cores with a Geoprobe sampler having the ability to drill or push through the concrete and brick of the historic fill, then used a combination of hydraulic push and high-frequency vibration to sample subsurface layers to a depth of 36 ft (11.0 m), raising continuous cores contained within plastic sleeves in five foot segments. An additional seven cores were taken from utility alignment trenches. Using historic evidence (Love, 1935), and working around the buried foundations of 18th and 19th century buildings buried in the fill, we arranged our cores in an L-shaped array to capture variability in both downriver (north-to-south; F15–F6 and R3–R6) and transverse (east-to-west; F16–F9) directions, with special attention to a "backswamp" indicated on

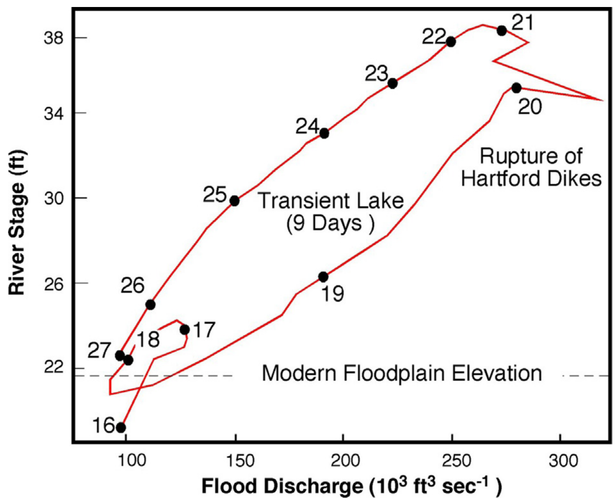

Fig. 3. Plot of river stage and flood discharge at Hartford, Connecticut for the 1936 flood of record. Redrawn from Kinnison et al. (1938). Numbered data points are the mean values for that day in March, 1936. Hysteresis (loop curve) indicates hydraulic ponding. Metric conversions are 0.305 m ft^{-1} and 0.028 m^3 ft^{-3}.

early historic maps that had been obliterated by later development (S1–S5). In all, we examined 47 subsurface records, consisting of sixteen engineering boreholes and 31 continuous cores.

The resulting stratigraphy (Figs. 5 and 6) was based primarily on detailed visual descriptions of the Geoprobe cores in the laboratory. Cores up to 11 m long were: slit open; split longitudinally; placed under a strong light source; scraped, wetted or dried to enhance textural (grain size) and color (Munsell) contrasts; subdivided into natural lithologic units; and carefully measured. Description protocols were systematized with a consistent set of unit designations after independent description of five reference cores (S3, P2, R5, R3, F9, F6, and F15). Unit elevations were based on original depths below surface, which were reconstructed by multiplying the percent compaction within core segments by the percent length of the unit within the segment. Samples for radiocarbon dating, pollen, granulometry, and visual description of the fine-sand fraction were rectified relative to surface elevations as well. Loss on ignition was not done because the sediment was chiefly inorganic clay and silt.

Pollen analysis was carried out under contract to consulting palynologist Linda Scott Cummings, based on blind samples sent to her by mail. Though conventional pollen diagrams were drawn, the quality and reliability of the pollen stratigraphy was greatly compromised by the high mineral content of the sediment, low pollen counts caused by poor preservation in oxidizing soils, bioturbation, and an inability to differentiate modern pollen (reaching the site by direct airfall or alluvial transport) from relict pollen being reworked alluvially from previous sediment storage. Between March 21 and October 31, 2001 eleven samples were taken from the Connecticut River to assess pollen concentration in the river water. Granulometry was done on 42 samples averaging 50 g each (larger samples would have required a less precise sampling protocol) from Core R3 taken between −1.5 and +6.5 m elevation by wet sieving at 4 ϕ (0.0625 mm), followed by pipette analysis for the silt and clay fractions and dry sieving the sand fraction. Samples were dispersed with 5 g sodium phosphate (Calgon) in an ultrasonic bath. After wet sieving, the very fine sand fraction was isolated and examined for notable changes in properties up the core. AMS (accelerator mass spectrometry) radiocarbon dating was done on samples of charcoal, wood, and leaf litter by Beta Analytic, Inc. All samples were carefully handled, pretreated, corrected for isotopic fractionation, and reported in uncalibrated years BP. Bulk magnetic susceptibility was performed for the reference core (R3) by passing it through an induction coil and taking measurements at 2 cm intervals using a Bartington MS2 system.

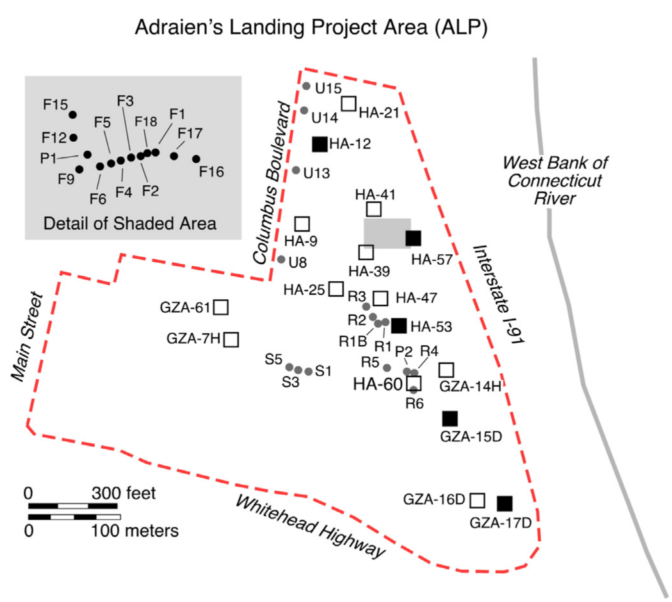

Fig. 4. Location of engineering borings (squares) and Geoprobe cores (circles) within the project area (dashed red line) relative to the west bank of the Connecticut River and key roads in downtown Hartford, Connecticut. Large gray box outside the project area is a blow-up of core locations for the smaller gray box within it. Solid boxes show borings from which pollen and radiocarbon samples were taken. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

6. Results and interpretations

6.1. Core description

Examination, comparison, and correlation of all cores allowed us to recognize discrete lithostratigraphic units, most of which were consistent in position from core to core, with minor variations in detail. Below is a general description and interpretation for each unit in the study area, beginning with the lowest (oldest).

6.1.1. Unit 1 – diamict

Compact and hematite-stained diamict was penetrated in the four deepest engineering boreholes below −6 m elevation. It was dominated by crushed shale and arkosic sandstone, but also contained rounded and freshly-fractured pebbles of crystalline metamorphic rock from the catchment highlands. This unit is interpreted as lodgment till, based on degree of compaction, shear fabric, and bulk strength. Some cores exhibit oxide staining and disaggregation along fractures.

6.1.2. Unit 2 – red silt and clay

Compact, hematite stained, massive to laminated clay and silt containing varves at the mm and cm scales characteristic of the Hartford area. Unit interpreted as the bottomset (lakebed) facies of Glacial Lake Hitchcock.

6.1.3. Unit 3 – sand & gravel

Beneath an elevation of −2.58 m, the deepest and most westerly Geoprobe core (F15) contained a matrix-supported, graded-bedded, sandy gravel beneath a cleanly washed, medium- to

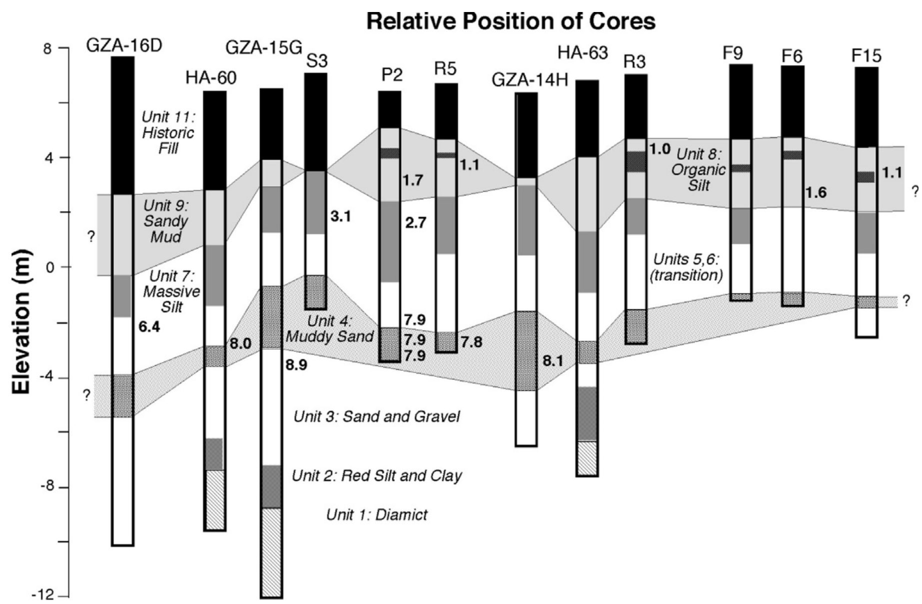

Fig. 5. Panel (fence diagram) of borings (located on Fig. 4) showing vertical positions of key stratigraphic units (named on diagram), described in text, and drawn schematically as Fig. 6. Relative position is parallel to the river in a northwesterly direction. Boldface numbers indicate radiocarbon dates in thousands of calendar years before present, as listed in Table 1. Two envelopes showing correlations are indicated.

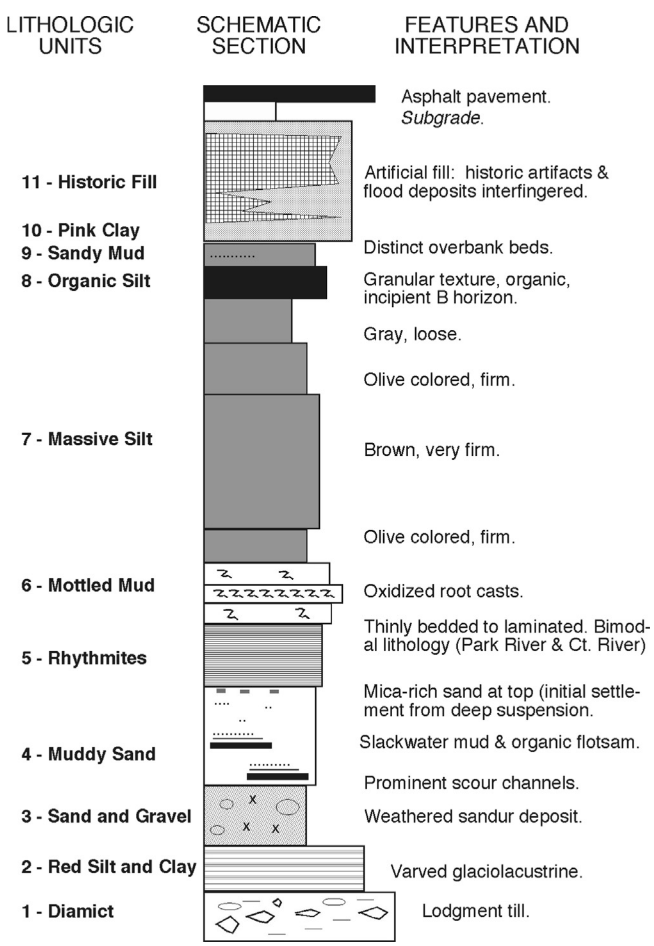

Fig. 6. Generalized stratigraphy for all boreholes and Geoprobe cores showing the numbers and names of key stratigraphic units, and their approximate relative thicknesses. Refer to Figs. 4 and 5 for locations and elevations. The column labeled features and interpretation provides a brief description and (or) interpretation from the text.

coarse-grained, sheet-bedded sand with rare inclined bedding. The entire unit was oxidized yellowish brown. A distinct patina (black) was restricted to the upper surface of the arkosic clasts, and is interpreted to indicate downward infiltration of vadose water. Similar material was widely present but poorly recovered during split-spoon sampling, owing to lack of cohesion. Based on measured elevations in five cores, there is up to 4 m of relief on both the lower and upper contacts of this unit. Unit 3 is interpreted to indicate *in situ* weathering of sediment deposited by fluvial traction on a braided gravelly surface that was locally channeled after being weathered.

6.1.4. Unit 4 — muddy sand

All of the cores penetrated highly variable basal units dominated by gray, banded, muddy, fine sand that was devoid of pebbles. Bedding is generally indistinct, with massive muddy sand grading upward to horizons of cleanly washed sand on scoured contacts. The deepest cores penetrated thick sequences of rhythmically bedded slackwater sediments, each with a sequence of washed fine sand overlain by mud, clay, and plant macrofossils including large chunks of bark-covered wood up to 8 cm thick. For example, Core R5 contained 15 conformable rhythmite sequences within in a 42.8 cm thick interval. Some sand beds contain lenses of matrix-supported red-shale chips (intraclasts) similar to those present as clasts in the underlying till. In every core, the uppermost bed was conspicuously rich in mica. The bulk of this unit is considered a product of pulses of rapid deposition by turbid basal flows and sediment-charged suspensions. The high core-to-core variability, and frequent evidence of channel scour indicate highly variable conditions of transient erosion followed by surges of sediment in temporary suspension. The consistent, mica-rich interval at the top may represent the initial episode of protracted deep submergence.

6.1.5. Unit 5 — rhythmites

Lying conformably above the mica-rich sand in nearly all of the cores is a sequence of color-banded mud with a bulk texture dominated by silt. Each rhythmite exhibits from the base up a sequence of: muddy fine to very fine sand characterized by upriver (catchment) lithology and slightly elutriated at the base; light-brown or pink silty clay from a Jurassic red-bed (hematite-stained) source; and dark beds containing a mixture of settled clay, organic material and/or oxide staining. Though highly variable, rhythmites average about 1.5 cm in thickness. Mottles are present throughout, increasing in concentration upward as the rhythmites become less distinct. Unit is interpreted as a series of dozens of submergence events on a flattening surface in which bioturbation, chiefly by plant roots, increases upward. The consistent pattern of fining upward within individual rhythmites, combined with the color transition (light-pink-dark), indicates: an early flush of suspended sediment being carried by the main flow from the Connecticut River; a deeper inundation accompanied by a flush of suspended sediment associated with the Park River, an accumulation of slackwater clay; drainage; and an interval of incipient subaerial weathering after each rhythmite event.

6.1.6. Unit 6 — Mottled Mud

Every core exhibits several decimeters of massive mud with a bulk texture generally similar to that of the underlying rhythmites, but with a lower sand content. Clay- and oxide-rich root casts and small concretions resembling peds are common, some of which appear to have been re-transported into incipient lag horizons. Typically, the concentration of root casts increases upward to a maximum before decreasing. Unit 6 appears to be a product of continuing slackwater sedimentation on an increasingly vegetated surface that is responsible for the turbation. The zone of maximum concentration of root casts is interpreted as a nascent soil beneath a vegetated surface, an interpretation consistent with the absence of this unit in Core S3, which is located within a backswamp on historic maps.

6.1.7. Unit 7 — Massive Silt

Unbedded, clay-rich, sand-poor, silt overlies a fairly abrupt, gradational, conformable contact. Although the color varies between and within cores, the general sequence is from gray near the base to grayish brown in the bulk of the unit. The upper third of the unit is conspicuously less brown, grading first into an equally stiff olive-colored sediment, then into a notably softer, sandier, and grayer silt. This unit appears to have formed as a result of a prolonged period of slackwater deposition on a continuously vegetated surface capable of destroying all bedding and homogenizing the sediment. The brown, olive, and gray color sequence upward indicates a diminishing degree of subaerial oxidation, likely a surrogate for sedimentation rate.

6.1.8. Unit 8 — Organic Silt

The gray, soft, uncompacted, massive sandy silt of the underlying unit continues upward, but with a conspicuous color change to nearly black, due to a rising concentration of decomposed humus. This unit also contains discrete grains of charcoal, an incipient granular blocky soil structure, oxidized ped surfaces, and rare root casts, and has a higher bulk strength than sediments above and below. In Core S3 the unit is exceptionally dark and contains a few arkose pebbles (manuports?) worked down from above. Core F9 exhibits a series of discrete black bands in an otherwise weakly bedded fine sand. Core R3 exhibits a massive interval with a conspicuous blocky ped structure. Unit is best expressed (Cores R5 and R3) where it is topographically highest, perhaps on the crest of a broad stationary levee. This is based on the progressively higher elevation of all units above Unit 4 (Muddy Sand) in this vicinity. This unit is interpreted as indicating a marked reduction in sedimentation rate and conditions more favorable to soil formation.

6.1.9. Unit 9 — sandy mud

Lying conformably above the organic silt and incipient soil is an interval of sandy mud that is conspicuously less organic and much coarser-textured. It is also weakly bedded, especially near the top, with visible stringers of fine sand and silt. Laminae exhibit load structures, vertical offsets, and truncations beneath the overlying fill. Sediment comprising this unit is interpreted as having been deposited by flood flows from suspension with possible weak traction bedding during flood surges. Preservation of bedding is an artifact of having been isolated from bioturbation by emplacement of the overlying fill.

6.1.10. Unit 10 — pink clay

A conspicuous deposit of pink clay lies beneath, is interbedded with, and overlies historic fill in most cores. It is similar in hue to the pink component of the rhythmites, but is more clay-rich, more cohesive, and brighter in chroma than anything above the glacial deposits. Where undisturbed, the pink clay exhibits delicately banded rhythmites of gray silt—clay overlain by pink clay. This unit is interpreted as a slackwater deposit from a Park River sediment source.

6.1.11. Unit 11 — fill

Although heterogeneous and highly variable core-to core, the general sequence from the base upward consists of: clumps of sandy mud similar in texture to the underlying alluvium mingled with decomposed woody debris along sharp intraformational contacts; demolition debris, identified by brick fragments, broken glass, and charcoal; concentrations and layers of coal ash and clinker; and finally by gravel, which is dominated by crushed traprock and to a lesser extent, locally arkose. Just below the 20th century asphalt pavements is the modern subgrade, which we did not investigate. This unit is interpreted to show a common sequence of historic modifications involving: re-grading, scraping, and dumping; the abandonment and demolition of early buildings; industrial activities; and re-grading for pavement installation. All of these human-emplaced deposits were intermittently flooded, as indicated by beds of slackwater sediment.

6.2. Granulometry

The previous interpretations based on sedimentary structures, weathering, and bulk appearance are independently supported by an analysis of sediment texture (Fig. 7). The basal Muddy Sand (Unit 4), which contains evidence for flood surges and local scour, is by far the most sandy unit of the Holocene floodplain sequence, consisting of up to 95% sand. This is consistent with the interpretation that Unit 4 is a product of local scour, highly variable conditions, and deposition by turbulent flows associated with basal traction. This texture, interpreted as having been created under shallow overbank flows, abruptly gives way to the dominance of silt with at least 10% clay in the rhythmite sequence (Unit 5). This texture, which indicates slackwater sedimentation, continues upward into the Mottled Mud (Unit 6).

The interval spanning more than a meter of sedimentation from −0.5 to +0.5 m elevation exhibits an unusual change in the sand texture. While the total amount of sand ($<3\ \phi$) relative to the clay and silt fractions falls from about 18% to about 9%, the proportion of sand coarser than 3 ϕ (fine, medium, and coarse) increased from 0 to 8%. This coarsening of the sand fraction at a time when the bulk sediment was fining upward is likely due to a slight increase in stream power during the rising limb of the flood hydrograph prior to a progressively longer periods of slackwater submergence. Stratigraphically, this coincides precisely with the rapid upward growth of the stationary levee and a steady increase in bioturbation.

The Massive Silt (Unit 7) constitutes about half the vertical interval in most cores. It is also the most texturally uniform unit, an observation consistent with the evidence for deep bioturbation and oxidation weathering.

Two features of the core are consistent with requirement that the volume of the reservoir was diminishing as the floodplain aggraded and its elevation increased. Clay content dropped from about 30% to nearly zero percent as the bottom of the transient reservoir (the top of the floodplain) rose upward by more than four meters from an elevation of 0.5–4.6 m. Clay had more opportunity to settle when the volume was larger because the retention time of the temporary reservoir was longer and the downstream flux of floodwaters consequently slower. The same interpretation is suggested by the seemingly exponential increase in the fine sand content in the upper part of the Massive Silt (Unit 7, 2.5–4.2 m elevation) and the corresponding decrease in silt content over the same interval. Rising transport power is indicated, even though the system remains generally below that of active traction.

The highest clay content within the core occurs within the historic Pink Clay (Unit 10): an abundance nearly twice that of any other sample. This reflects deep slackwater settling at a time when the Park River was the dominant sediment source, perhaps when engineered dikes were preventing sedimentation from a Connecticut River source. Some time after deposition of the pink clay, the Park River was channelized underground beneath the levee to be pumped up to the opposite side of the dikes.

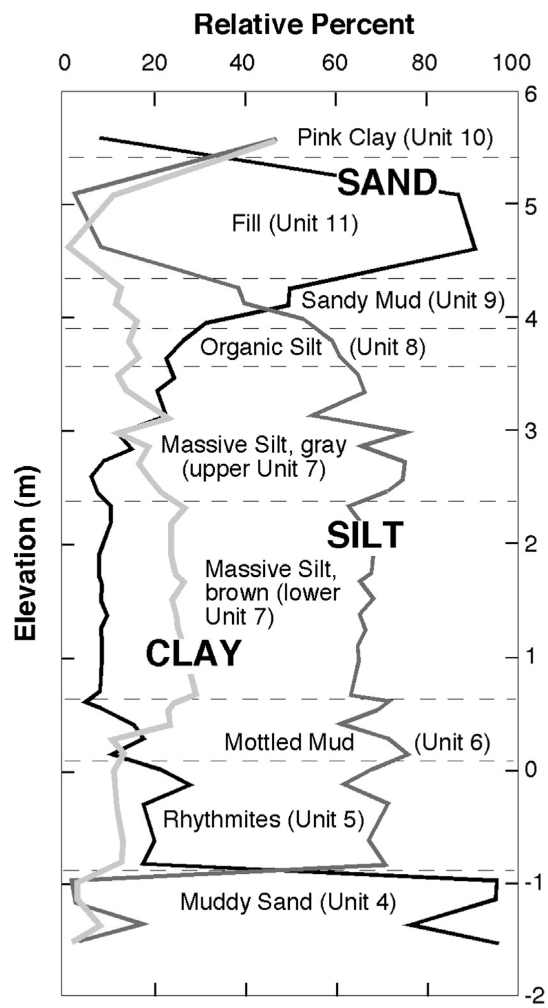

Fig. 7. Granulometry of 42 samples from Core R3 showing relative percent sand (black), silt (dark gray) and clay (light gray). Dashed horizontal lines show boundaries between units (Fig. 6) in the sampled core.

6.3. Magnetics

Magnetic susceptibility measurements were made only in the centrally located and representative core R3. Consistent values of mass magnetic susceptibility (χ_m) for prehistoric units ranged from <5 cm^3 g^{-1} in the brown middle member of the Massive Silt at 1.5 m elevation to 155 cm^3 g^{-1} in the soft sandy mud within the same unit but directly beneath the base of the organic silt and incipient soil just above 3 m elevation. This relationship suggests that susceptibility is measuring the degree to which ferrous iron on mineral grains remains unoxidized. Spikes of susceptibility elsewhere in the core coincide with sand content, which is often the norm. Individually targeted samples confirmed that magnetic susceptibility was not an indicator for the degree of hematite stain. Dramatic rises in susceptibility by more than an order of magnitude occurred in fill dominated by coal residues.

6.4. Pollen

Pollen analysis was reported separately (Cummings and Moutoux, 1999). The split spoon samples could not be used to reconstruct a continuous sequence and were found to be contaminated with surface pollen. Core P2 was collected specifically for pollen analysis and yielded a low-resolution pollen sequence broadly consistent with, but of poorer quality than typical for cores from isolated ponds. In addition to low concentrations of pollen grains, the most basic methodological problem was the inability to differentiate modern pollen deposited directly by airfall from relict and modern pollen brought in by river transport. For example, the Connecticut River water samples yielded between 100 and 2000 pollen grains per liter. Comparison of the identified pollen grains with the date of sample collection suggests "some of the pollen types appear to be reintroduced into the river after their pollinating season" (Cummings and Moutoux, 1999). Although the pollen identified within the water samples is consistent with the modern flora of the Hartford region, we were unable to exclude the effects of upriver pollen transport.

Within the sediment cores, the extraordinary range of pollen grain density per sample, ranging from 0 to 5602 grains cm^{-3}, was due to the combined effects of the original deposition facies and, more importantly, the relative degree of decomposition within the drained and turbated floodplain soils between deposition events. High pollen concentrations occur only where a black organic surface was abruptly covered by a sediment unit thick enough to isolate it from the weathering zone, for example in flotsam from the Muddy Sand (Unit 4), the Organic Silt (Unit 8) and within the historic organic stringers of Unit 9 (Sandy Mud) directly beneath thick fill. In each case, the pollen concentrations diminished exponentially with depth due to decomposition.

Pollen was essentially absent from mineral sediment below the onset of hydraulic ponding (Unit 5). The abrupt shift toward accumulation of fine-grained, massive, bioturbated, oxidized floodplain silt (Unit 6–7 contact) coincides with an abrupt drop in the degree of pollen preservation. Seven small spikes of Artemisia (ragweed), an indicator of dry, disturbed surfaces, occur within the floodplain sequence beneath the middle of the Massive Silt (Unit 7). The presence of *Plantago sp*. (plantain), *Zea mays* (corn), and *Cirsium* (thistle) just below the Organic Silt (Unit 8) suggests that Native Americans were clearing and cultivating the landscape at or upstream from the coring locality.

6.5. Radiocarbon dating

The sixteen AMS ages reveal a concordant series between 8900 ± 40 and 990 ± 40 BP (conventional radiocarbon years BP; 10,030–930 calibrated years age BP; 8080 B.C.E. to 1020 A.D.) that was consistent with the order of stratigraphic units (Table 1). There were no evident outliers or inversions in the dates. Variation in age within units is likely due to differences in stratigraphic depths that could not be resolved due to differential elevations. The relative uniformity in the ^{13}C fractionation, the dearth of carbonate sources, the evidence for short carbon residence time in the soils, and the fact that the radiocarbon ages are in correct stratigraphic order indicate that the numerical chronology is reliable.

6.6. Floodplain stages

The lithologic, magnetic, pollen, and chronostratigraphy of the ALP study area indicates four major stages of floodplain development. Though of unequal duration, each was bracketed by the river's response to a physical threshold driven largely by the flow of water through the Straits. These stages are given local names, each in italics.

The *Dry Terrace Stage* (>10.0–13.0 calibrated ka) was a protracted interval of freely drained conditions during which the braided stream deposits associated with the downcutting of Glacial Lake Hitchcock stood above the reach of an entrenched Connecticut River. A hydraulic dam may have been present, but was not sufficient to raise the back-flood to the level of Adriaen's Landing. Floods rose and fell within a quasi-equilibrium channel whose cross-sectional area was set by the hydraulic geometry (channel area, bed slope, bed and bank materials, sinuosity), flow velocity, and stage-discharge frequencies. This channel and its adjacent floodplain were aggrading below Adriaen's Landing.

The *Floodway Stage* (10.0–8.6 calibrated ka) began with the first, local overtopping of the Pleistocene terrace. This was as a consequence of the inevitable rise of the thalweg associated with a gently sloping longitudinal river profile graded to a slowly rising sea level. Hydraulic ponding was present at the Straits (Middletown), but its upper limit had not yet reached the ALP. Hence, the flood mechanism was that of a normal floodplain, with floodwaters overtopping channel banks and spilling outward and forward over the former terrace. When this stage was reached, the surface at Adriaen's Landing would have been forested, probably by a co-dominant mixture of pine and oak, based on the regional pollen record. The erosion and channeling associated with the first few overflows were dramatic, creating nearly 4 m of vertical relief on the former terrace surface (Fig. 5). Eventually, the highly variable conditions of scour and rapid deposition during the rising limb of the hydrograph gave way to backflooding during near-peak and recession flood flows. Some combination of multiple flood surges from the main channel during a single event and closely spaced seasonal floods were responsible for the aggradation of the sandy, twiggy, and sharply bedded flotsam deposits within former erosional channels (Fig. 6).

The *Transition Stage* (±8.6–7.3 calibrated ka) began when, for the first time, the channel was overtopped by the hydraulic back-flood mechanism before it could spill out from the floodwave. Fluvial erosion and sediment processes shifted away from traction (governed by basal shear stress) to those dominated by suspension (Stokes settling). Stratigraphically, this transition coincides with the sole conspicuous concentration of oversized mica flakes in the stratigraphy, indicating clarification of the water column before complete drainage (Fig. 6). This horizon lies directly beneath the basal rhythmite containing pink silt drawn from the Park River source area. A switchover to the Park River source requires that the main current in the Connecticut River be stalled, or that some mechanism bring the red suspended sediment northward approximately half a kilometer. Because flow in the Connecticut River is colder than that of the Park River, a thermally driven convective overflow mechanism might have operated during spring floods. Alternatively, the required northward flow of red silty clay might be due to a large stationary back eddy.

Deposition of 1–2 m of rhythmites by slackwater ponding would have elevated the top of the river bank, enlarging the channel cross section relative to that of the preceding Floodway Stage when flooding was controlled by the downriver-propagating, rather than upriver-propagating processes. Bank overflows ceased. The upstream limit of the zone of hydraulic ponding migrated upriver toward Windsor, diminishing the sediment supply near Hartford, reducing the sedimentation rate. This likely explains both the reduction in sand content from about 20% to about 5%, and the increase in bioturbation indicated by the shift from rhythmically bedded to massive sediment, respectively (Fig. 7). A vegetated floodplain above an underfit channel was now in place.

The *Stable Floodplain* Stage (7.3–0.4 calibrated ka) was the longest and most stable interval of the Holocene history. A regular back-flood regime was in place. The floodplain surface (also the base of the transient pond) rose slowly (0.9 mm/yr) at a rate about half that of sea level (1.7 mm/yr). Nevertheless, this was faster than the rise in peak stage (0.6 mm/yr; based on the Cone equation, and assuming a triangular cross section). The consequence of this was a reduction in the volume of the reservoir, which, in turn, required a reduction in flood retention time, which, in turn, required the average flow-through velocity to increase. The gradual and exponential coarsening upward of the slackwater floodplain sequence under conditions of hydraulic ponding was the result.

Any discussion of the *Historic Stage* (0.4 calibrated ka to present) must begin with the contact between the Sandy Mud (Unit 9; laid down before urbanization), and the overlying Historic Fill (Unit 11), which is bracketed by, and intercalated with, beds and deformed stringers of the Pink Clay (Unit 10; Fig. 6). Within the Sandy Mud, an abrupt coarsening of bulk texture, the preservation of soft plant tissue, discrete (unturbated) beds of flood-deposited sand, and evidence of loading of sedimentary contacts (soft-sediment deformation) all indicate a significant jump in sedimentation rate prior to significant urbanization. These multiple lines of evidence suggest that increased runoff from watersheds in the process of being cleared for farms during the 18th and early 19th centuries was likely involved. The abundance and purity of the pink clay is restricted to the time of floodplain engineering, presumably after the main flow of the Connecticut River had been shunted southward by dikes, but before the Park River was diverted underground and pumped up and over the dikes.

7. Geoarchaeology

7.1. Historic site

Adriaen's Landing is an important historic archaeological site. Although artificial fill dominates the upper 2–4 m of the floodplain by mass, it is bracketed and interbedded with naturally formed, in-situ geologic units, most conspicuously Unit 11, the Pink Clay. This places the material remains of the 17th–18th century Euro-Americans – complete with the foundations of abandoned buildings – into a stratified archaeological context not unlike that of a prehistoric midden bracketed by estuarine sediments. Core F9 contained a unique bed of very dark organic material within mineral mud containing broken shell fragments as well as bone with saw marks. We interpret this either as dump or as a garden plot in which food refuse was deliberately added to enrich the soil. A heavy concentration of ceramic fragments in Core S3 was dispersed within the upper part of the floodplain mud in a locality adjacent to a pottery manufacturer and within the back-swamp indicated by historic mapping. Near the base of the fill was a slab of soapstone drilled through during coring. Core F15 contained layers of shell fragments.

Prior to the arrival of Europeans, the ALP area would have been unattractive for residential settlement. In 1790, it lay at least 91 m (300 ft) west of a river bank at least 10 m high that was being scoured by annual floods. During our coring operation, we found only one potential prehistoric artifact, a very smooth pebble of crystalline rock, which plausibly could have been dropped or thrown by human beings or some other animal agency. An ice-rafting or driftwood-rafting agency is inconsistent with the complete absence of any sediment coarser than medium sand within the main floodplain sequence.

7.2. Prehistoric sites

The distribution of prehistoric sites within the CRAL in space and time has been interpreted largely on the basis of inferred ecological requirements (Nicholas, 1988; Dewar and McBride, 1992; Curran, 2003) and more objective geographic patterns (Forrest et al., 2006, Appendix II, Regional GIS Archaeological Model). Here, we correlate site distributions with alluvial reaches and sectors within them, based on site files continuously updated by the Office of the State Archaeologist. We note that the densest concentration of prehistoric sites within the CRAL ($n = 18$) occurs as a cluster within a veneer of overbank deposits on a low terrace tread above the Connecticut River at the boundary between the freely meandering Glastonbury Reach to the north and the confined Rocky Hill Reach to the south (small box on Fig. 1). As with the ALP site, this setting has the advantages of free drainage during most of the year, frequent burial by transient lacustrine slackwater sediments, no significant erosion, and proximity to the Connecticut River. In other words, both sites have the taphonomic advantages of floodplain sedimentation without the disadvantages of erosion or channel migration. The concentration of sites has the additional advantage of overlooking a very important ecological boundary separating the riparian ecosystem of the Glastonbury meanders (to the north) from the more constricted "canyon-like" reaches to the south. A former rapids was likely present at this site earlier in the Holocene because of the abrupt increase in stream power, perhaps concentrating human activities.

Another concentration of sites follows the edge of the high terrace overlooking the floodplain within the Glastonbury Reach. There, eighteen Late Archaic sites overlook the two most prominent meanders, which today are trimming the edge of the terrace, maintaining unvegetated slopes that can be used as a convenient access up the bluff in a setting where bluff-top eolian deposition (Thorson and Tryon, 2003) facilitates preservation. No sites are mapped on the floodplain below the terrace here, presumably because the surface is much younger and the channel is actively migrating.

The density of prehistoric sites falls off dramatically further north within the Glastonbury and Hartford reaches, where the river is more distant from flanking terraces, and where those terraces have not been cut by bank erosion during historic time. The density of sites picks up again within the Windsor Reach. Most archaeological sites there are located on elevated portions of the floodplain (levees) in a setting where the river has shifted laterally by avulsion, but has not meandered, thereby maintaining preservation.

7.3. Archaeological model

The most significant result of our study is the creation of a mechanistic sedimentary model for the Holocene archaeology of the Glastonbury, Hartford, and Windsor reaches that can be independently compared with models based on paleoclimate, cultural chronology, and human ecology.

The *Dry Terrace* Stage of the alluvial chronology is roughly coincident with the PaleoIndian Period (11.0–9.5 ka). Within the CRAL, the Connecticut River was aggrading its channel within an entrenched valley, the outer boundaries of which approximate that of the present floodplain. The lower thalweg elevation might have exposed bedrock falls and rapids that would have promoted fishing in the Enfield and Windsor reaches. Our evidence from the ALP indicates that much of the terrace near the river was scoured by overbank flooding during the subsequent Floodway Stage, likely leading to the destruction of most sites that had existed along the bank.

The *Floodway Stage* and the *Transition Stage* were time-transgressive, beginning first in the Cromwell reach and ending at Enfield where the effects of hydraulic damming disappeared. Hartford experienced the Floodway Stage between 8.9 and 7.8 ka, an interval roughly equivalent to the Early Archaic (9.5–8.0 ka) and the beginning of the Middle Archaic (8.0–6.0 ka) Periods. It experienced the Transition Stage ($\sim$6.4 ka) during the last half of the Middle Archaic (8.0–6.0 ka). During the temporal shift between these stages a physical geographic band (ecotone) perhaps a few km wide may have provided special resources. This zone may have resembled the shoreline of a muddy lake that varies in stage. It would have graded southward into a broad, stable, vegetated, zone where back-flooding happened at least every few years, and northward to a narrower, more volatile floodplain setting where overbank flooding in a rising channel was the norm. Contemporaneous occupation sites would have been slowly buried and bioturbated to the south, rapidly buried within the band, and eroded to the north.

The longest and most predictable alluvial stage was the *Stable Floodplain* Stage, which began very close to 6.4 ka and which experienced little change until some time after 2.7 but before 1.7 ka. These dates correspond almost perfectly to the Late Archaic Period (6.0–2.7 ka), the first time in the CRAL when sites become abundant and well preserved. Recurrent floods were deep enough to deposit clay and silt on the edges of terraces, but drainage conditions were good enough to allow luxurious growth of vegetation and the development of oxidized floodplain soils.

The Early Woodland Period (2.7–2.0 ka) and Middle Woodland Period (2.0–1.0 ka) were times of very few sites in the CRAL, a pattern that has been observed regionally (Fiedel, 2001). These archaeological periods correspond to the gradual reversal of the stratigraphic sequence at Hartford: away from slower rates of deposition, finer grained materials, more intense weathering, and deeper oxidation that characterized the first two thirds of the Stable Floodplain Stage, and which gave rise to the brown floodplain silts dominating the central parts of every core examined. The subsequent interval between about 2-3 ka and 1 ka, also within the Stable Floodplain Stage, was characterized first by a return to less well drained conditions leading to olive colored floodplain deposits, and then much looser, grayer alluvium indicating more rapid deposition of coarser materials. This change was probably fairly extensive within the CRAL because, by this time, the retention time of the reservoir was diminishing. A more volatile flood regime may have hindered human settlement on the floodplain.

The Late Woodland Period (1.0–0.4 ka) was a time of landscape stability, soil formation, intense occupation within the CRAL, and a shift towards increased use of pottery and cultigens, larger populations, food storage, and more sedentary communities. Stratigraphically, this interval coincides with the Organic Silt, which was observed within every Geoprobe Core as a concentration of humus and the incipient formation of a weak B horizon. Climatically, Late Woodland occupation and this incipient soil partially coincide with an interval of northern hemispheric warmth known as the Medieval Climatic Optimum. This period of warmth and reduced flooding may have also promoted the use of cultigens in the CRAL. The return to coarser, less weathered, and less compacted sediments suggestive of a more volatile flooding regime may be associated with the subsequent Little Ice Age climate (1300–1850 AD), though explanations based on sea level and forest clearing cannot be ruled out.

8. Conclusion

We see two main contributions of this article. Scientifically, it presents a conceptual model for alluvial stratigraphy of floodplain settings where long-term hydraulic ponding occurs. At Adriaen's Landing, channel and levee deposition was followed not by a series of overbank floods *per se*, but by a series of progressively more transient lacustrine inundations spanning a 10,000-year period, each of which migrated upvalley from the site of ponding. Over time, and as vertical sediment accretion from suspension occurred, the accommodation space below peak flood stage was reduced, forcing a shortening of the flood residence time and a consequent coarsening of the bulk texture. This same scenario can be applied elsewhere to large river systems in tectonically stable contexts wherever valley constrictions cause significant hydraulic ponding.

From the perspective of cultural resource management (CRM), the Adriaen's Landing Project provides an example of an alternative archaeological research goal. There, the effort was targeted not toward the recovery of archaeological objects *per se*, but to the recovery of stratigraphic data that could be used to build a geoarchaeological model, into which known objects and future sites could be plugged.

Acknowledgments

Funding for this project was provided by a contract to Archaeological and Historic Services (AHS) from the Connecticut State Office of Policy and Management. Mary Harper and Phil McClellan managed this project on their respective sides. Boreholes were located based on historic cartographic analysis conducted by Dr. Michael Raber of Raber Associates. Dr. Linda Scott Cummings of the Paleoresearch Institute performed the palynological and phytolith analysis. James A. Hyatt assisted with magnetic susceptibility measurements. Our field crews performed admirably in often difficult and occasionally hazardous conditions, with special thanks to James Poetzinger, Eric Pomo, Karen Morando, Tim Kuskowski, Brian Garro, Lisa Centola, Timothy Ives and Ross Harper. Anonymous reviews greatly improved this manuscript.

References

Ahearn, E.A., 2005. Estimates of the Magnitude and Frequency of Flood Flows in the Connecticut River in Connecticut. Open File Report 2005-1369, U.S. Geological Survey, Washington, D.C., 12 pp.

Bierman, P., Lini, A., Zehfull, P., Church, A., Davis, P.T., Southon, J., Baldwin, L., 1997. Postglacial ponds and alluvial fans: recorders of Holocene landscape history. GSA Today 7, 1–8.

Brakenridge, G.R., 1988. River flood regime and floodplain stratigraphy. In: Baker, V.C., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. Wiley, New York, pp. 139–156.

Briffa, K.R., Osborn, T.J., Schweingruber, F.H., Harris, I.C., Jones, P.D., Shiyatov, S.G., Vaganov, E.A., 2001. Low-frequency temperature variations from a northern tree-ring density network. Journal of Geophysical Research 106, 2929–2941.

COHMAP Members, 1988. Climatic changes of the last 18,000 years: observations and model simulations. Science 241, 1043–1052.

Cummings, L.S., Moutoux, T.E., 1999. Stratigraphic Pollen and Phytolith Analysis, Adriaen's Landing, Connecticut. Technical Report 99-48, prepared for PAST, Inc., Paleo Research Labratories, Golden, CO.

Curran, Kathryn, 2003. Geochronology from archaeology: an example from the Connecticut River valley. In: Cremeens, D.L., Hart, J.P. (Eds.), Geoarchaeology of Landscapes in the Glaciated Northeast, Bulletin 497. New York State Museum, Albany, New York, pp. 151–162.

Dewar, R., McBride, K., 1992. Remnant settlement patterns. In: Rossignol, J., Wandsnider, L. (Eds.), Space, Time and Archaeological Landscapes. Plenum Press, New York, pp. 193–226.

Fiedel, S., 2001. What happened in the Early Woodland? Archaeology of Eastern North America 29, 101–142.

Forrest, D.T., Raber, M.S., Jones, B.D., Thorson, R.M., 2006. Archaeological and Historical Resource Study, Adriaen's Landing Project, Hartford, Connecticut. Report prepared by Archaeological & Historic Services, Inc., for the Connecticut Office of Policy and Management, Storrs, Connecticut, 100 pp.

Gayes, P.T., Bokuniewicz, H.J., 1991. Estuarine paleoshorelines in Long Island Sound, New York. Journal of Coastal Research (Special Issue No. 11), 39–54.

Gladfelter, B.G., 2001. Archaeological sediments in human alluvial environments. In: Stein, J.K., Farrand, W.R. (Eds.), Sediments in Archaeological Context. University of Utah Press, Salt Lake City, pp. 93–125.

Handman, E.H., Hildreth, C.T., 1972. Depth to Bedrock, Hartford North Quadrangle, Connecticut. In: Miscellaneous Investigations Series Map I-784-D. U.S. Geological Survey, Washington, D.C.

Houston, D.C., 1868. Colonel Houston's Report of the Survey of the Connecticut River. Report to the 40th Congress, 2nd Session, House Executive Document 153, U.S. Serial Set Vol. 1331, Washington, D.C., pp. 3–20.

Jahns, R.H., 1947. Geologic Features of the Connecticut Valley, Massachusetts as Related to Recent Floods. In: Water-Supply Paper 996. U. S. Geological Survey, Washington, D.C.

Kinnison, H.B., Conover, L.F., Bigwood, B.L., 1938. Stages and Flood Discharges of the Connecticut River at Hartford, Connecticut. In: Water-Supply Paper N836A. U. S. Geological Survey, Washington, D.C.

Knox, J.C., 1993. Large increases in flood magnitude in response to modest changes in climate. Nature 361, 430–432.

Lewis, R.S., Stone, J.R., 1991. Late Quaternary stratigraphy and depositional history of the Long Island Sound Basin; Connecticut and New York. Journal of Coastal Research (Special Issue No. 11), 1–23.

Love, W.D., 1974 [1935]. The Colonial History of Hartford, Gathered from Original Sources. Centinel Hill Press & Pequot Press, Inc., Hartford, Connecticut.

Metzler, K.J., Damman, A.W.H., 1985. Vegetation patterns in the Connecticut River flood plain in relation to frequency and duration of flooding. Le Naturaliste Canadien 112, 535–547.

Nicholas, G.P., 1988. Ecological Leveling. In: Nichols, G.P. (Ed.), Holocene Human Ecology in Northeastern North America: Interdisciplinary Contributions to Archaeology. Plenum Press, New York, pp. 257–296.

O'Connor, J.E., Webb, R.H., 1988. River flood regime and floodplain stratigraphy. In: Baker, V.C., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. Wiley, New York, pp. 393–418.

Patton, P.C., 1988. Geomorphic response of streams to floods in the glaciated terrain of southern New England. In: Baker, V.C., Kochel, R.C., Patton, P.C. (Eds.), Flood Geomorphology. Wiley, New York, pp. 261–277.

Patton, P.C., Horne, G.S., 1991. A submergence curve for the Connecticut River Estuary. Journal of Coastal Research 11, 181–196.

Pedereson, D.C., Peteet, D.M., Kurdyla, D., Guilderson, T., 2005. Medieval warming, Little Ice Age, and European impact on the environment during the last millennium in the lower Hudson Valley, New York, USA. Quaternary Research 63, 238–249.

Peltier, W.R., 1996. Mantle viscosity and ice-age ice sheet topography. Science 273, 1359–1364.

Peteet, D.M., Beh, M., Orr, C., Kurdyla, D., Nichols, J., Guilderson, T., 2012. Delayed deglaciation or extreme Arctic conditions 21-16 cal.kyr at southeastern Laurentide Ice Sheet Margin? Geophysical Research Letters 39 (L11706), 6.

Ridge, J.C., 2003. The last deglaciatoin of the northeastern United States: a combined varve, paleomagnetic, and calibrated ^{14}C chronology. In: Cremeens, D.L., Hart, J.P. (Eds.), Geoarchaeology of Landscapes in the Glaciated Northeast, Bulletin 497. New York State Museum, Albany, New York, pp. 15–45.

Rittenour, T.M., Brigham-Grette, J., Mann, M.E., 2000. El Nino-like climatic teleconnections in New England during the late Pleistocene. Science 288, 1039–1042.

Schuldenrein, J., 2003. Landscape change, human occupation, and archaeological site preservation at the glacial margin: geoarchaeological perspectives from the Sandts Eddy site (36Nm12), Middle Delaware valley, Pennsylvania. In: Cremeens, D.L., Hart, J.P. (Eds.), Geoarchaeology of Landscapes in the Glaciated Northeast, Bulletin 497. New York State Museum, Albany, New York, pp. 181–210.

Shaw, J., van de Plassche, 0, 1991. Palynology of late Wisconsin/early Holocene lake and marsh deposits, Hammock River Marsh, Connecticut. Journal of Coastal Research 11, 85–104.

Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., Miller, H.L. (Eds.), 2007. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, 2007. Cambridge University Press, Cambridge, UK.

Stone, J.R., Schafer, J.P., London, E.H., DiGiacomo-Cohen, M.L., Lewis, R.S., Thompson, W.B., 2005. Quaternary Geologic Map of Connecticut and Long Island Sound Basin. Map 2784, Scale 1:125,000. U.S. Geological Survey, Washington, D.C.

Thomson, J.R., Gannon, W.B., Thomas, M.P., Hayes, G.S., 1964. Historical Floods of New England. In: Water-Supply Paper 1779-M. U.S. Geological Survey, Washington, D.C., pp. 1–105

Thorson, R.M., Schile, C.A., 1995. Deglacial eolian regimes in New England. Geological Society of America Bulletin 107, 751–761.
Thorson, R.M., Tryon, C.A., 2003. Bluff top sand sheets in Northeastern archaeology: a physical transport model and application to the Neville site, Amoskeag Falls, New Hampshire. In: Cremeens, D.L., Hart, J.P. (Eds.), Geoarchaeology of Landscapes in the Glaciated Northeast, Bulletin 497. New York State Museum, Albany, New York, pp. 61–73.
Thorson, R.M., Webb, R.S., 1991. Postglacial history of a cedar swamp in southeastern Connecticut. Journal of Paleolimnology 6, 17–35.
Van de Plassche, O., 2000. North Atlantic climate-ocean variations and sea level in Long Island Sound, Connecticut, since 500 cal yr A.D. Quaternary Research 53, 89–97.
Varekamp, J.C., Thomas, E., Van de Plassche, O., 1992. Relative sea-level rise and climate change over the last 1500 years. Terra Nova 4, 293–304.
Webb, R.S., Anderson, K.H., Webb, T., 1993. Pollen response-surface estimates of late-Quaternary changes in the moisture balance of the northeastern United States. Quaternary Research 40, 213–227.
Wolman, M.G., Eiler, J.P., 1958. Reconnaissance study of erosion and deposition produced by the flood of August 1955 in Connecticut. Transactions of the American Geophysical Union 39, 1–14.
Wolman, M.G., Leopold, L.B., 1957. River Flood Plains: Some Observations on Their Formation. Professional Paper 282-C. U.S. Geological Survey, Washington, D.C., pp. 89–109