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Quaternary Geologic Map of Connecticut and Long Island Sound Basin

By Janet Radway Stone, John P. Schafer, Elizabeth Haley London, Mary L. DiGiacomo-Cohen, Ralph S. Lewis, and Woodrow B. Thompson

With a section on Sedimentary Facies and Morphosequences of Glacial Meltwater Deposits by Byron D. Stone and Janet Radway Stone

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Cover. Westerly view of the Hammonasset-Ledyard moraine as it disappears offshore beneath Long Island Sound at Meigs Point, Hammonasset Beach State Park in Madison, Conn. Photograph by Janet Radway Stone.

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Introduction

The Quaternary geologic map (sheet 1) and explanatory figures and cross sections (sheet 2) portray the geologic features formed in Connecticut during the Quaternary Period, which includes the Pleistocene (glacial) and Holocene (postglacial) Epochs. The Quaternary Period has been a time of development of many details of the landscape and of all the surficial deposits. At least twice in the late Pleistocene, continental ice sheets swept across Connecticut. Their effects are of pervasive importance to the present occupants of the land.

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

Connecticut is covered by 116 7.5-minute quadrangles, all available as U.S. Geological Survey 1:24,000-scale topographic maps with a 10-ft contour interval. Surficial geologic maps exist in various forms (either published, open-filed, or unpublished) for 98 of these quadrangles. We reviewed all 98 maps, and did reconnaissance mapping in the remaining quadrangles. An index map and a list of references to these quadrangle studies are included in Appendix 3. In the course of compiling this large body of data to create both the Surficial Materials Map and the Quaternary Geologic Map, we applied a consistent interpretive rationale; the result is that, in some cases, the original studies have been reworked or revised.

The contour intervals for the 1:100,000-scale base map are in metric units; however, all measurements for onshore units were made in English units based on the 1:24,000-scale quadrangle maps, where contour intervals are in feet. Therefore, dual measurements are provided for altitudes, distances, and thickness measurements in unit descriptions (see Description of Map Units). Altitudes and depths offshore are given in meters below mean sea level (MSL) for convenience in referencing the marine seismic-reflection profile data, which is recorded and described in metric units; the seismic-reflection profiles are the primary source of offshore geologic information.

Map Units

The map units are divided into three main groups: postglacial, glacial ice-laid, and glacial meltwater deposits. The postglacial deposits, formed by various processes after the recession of the last ice sheet, constitute ten map units. The glacial ice-laid deposits, poorly sorted materials deposited more or less directly by the ice sheets, constitute three units. The 13 glacial ice-laid and postglacial units are Statewide in distribution, except for such limitations as are imposed by the geologic processes involved; for example, coastal beach and dune deposits are, of course, restricted to the coast. The glacial meltwater deposits, laid down by the great volumes of meltwater produced during the shrinkage of the last ice sheet, include 6 Statewide depositional systems. The 6 systems of meltwater deposits are further differentiated into 204 units, closely restricted in geographic location and therefore also in age (see correlation chart on sheet 1); these units have been given informal names based on their geographic localities. In the Description of Map Units, the glacial meltwater map

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units are organized by type of depositional system within each basin and by their occurrence within 8 geologic basins across the State (figure 1 on sheet 2). The rationale for this profusion of units is based on their scientific importance; they provide detailed information on the mode of disappearance of the last ice sheet and the depositional processes operating around its margin. A comprehensive understanding of these units has many practical applications in studies of socio-economic importance such as ground-water availability and coarse aggregate resources.

Map Units Beneath Long Island Sound

Glacial and postglacial geologic deposits have long been studied on land in Connecticut. Similar deposits are also present beneath modern marine sediments in Long Island Sound; in fact, it is here that the thickest and most extensive Pleistocene deposits are found (figure 2 on sheet 2). Map units beneath Long Island Sound differ from those on land in two ways. First, offshore geologic units are mapped largely from analysis of seismic-reflection profile data instead of from direct observation. Secondly, due to the three-dimensional aspect of seismic-reflection data, map units in many places in the Long Island Sound basin are superposed and show vertical as well as areal distribution. Nearly all of the offshore geologic units of the Long Island Sound basin occur beneath a generally ubiquitous blanket of Holocene marine mud a few to 15 m (33 ft) in thickness which is not shown on the map (Lewis and DiGiacomo Cohen, 2000).

The distribution of late Quaternary geologic units beneath Long Island Sound was mapped from more than 4,000 line-km of high-resolution, seismic-reflection profiles (see figure 1 for locations) supplemented by vibracore data. These data were collected beginning in 1982 as part of an ongoing marine geologic mapping program conducted by the Connecticut Geological and Natural History Survey (CGNHS) in cooperation with the U.S. Geological Survey. Spacing between seismic lines that were collected perpendicular to the coast is about 1.6 km (1 mi) (locally 0.8 km (0.5 mi)), and about 4.4 km (2.7 mi) between tielines paralleling the coast. Vibracores (obtained through CGNHS cooperative programs with the U.S. Minerals Management Service) and submersible and remotely operated vehicle dives (made possible by cooperative programs with National Oceanic and Atmospheric Administration's National Undersea Research Center at the University of Connecticut) provided verification of near-surface geologic interpretations made from seismic-reflection profiles.

Varved lake clays and marine muds in the Long Island Sound basin have distinctive seismic signatures and were easily differentiated with minimal sampling (Lewis and Stone, 1991; Lewis and others, 1993; Lewis and DiGiacomo Cohen, 2000). Other geologic units (including moraine deposits, proximal and distal glaciolacustrine fan deposits, glaciodeltaic deposits, channel-fill deposits, and marine delta deposits) were mapped by analyzing internal-reflection characteristics of seismic units in a synergistic basinwide context. The offshore mapping was made possible by collaborative interpretation of seismic lines by glacial geologists familiar with the distribution and internal structure of terrestrial Quaternary deposits and by marine geologists familiar with the offshore seismic record. Systematic seismic-survey coverage of the entire basin (see figure 1) was also an important factor in the offshore mapping effort because knowledge of the areal distribution of seismic units is necessary to their interpretation as specific depositional facies and correlation with terrestrial geologic units.

Preglacial Landscape and Bedrock Source Areas

The bedrock beneath Connecticut consists of Precambrian and Paleozoic crystalline rocks in the eastern and western highlands (figure 3 on sheet 2); these rock units extend offshore beneath Long Island Sound. The Central Lowland of Connecticut is underlain by early Mesozoic arkosic sedimentary rocks locally intercalated with igneous basalt and diabase. Semiconsolidated sands, gravels, and clays of Cretaceous coastal plain strata underlie the southern part of the Long Island Sound basin and Long Island. The presence of coastal plain strata beneath Long Island has long been known from extensive water-well and test-boring data (Fuller, 1914; Suter and others, 1949); the extent of these units beneath Long Island Sound, terminated on its northern side by a fluvially carved, glacially scoured cuesta scarp, has only recently been mapped in detail from systematic seismic-reflection profiles in Long Island Sound (Lewis and Needell, 1987; Needell and others, 1987; Lewis and Stone, 1991).

Before glaciation, the crystalline and sedimentary rocks of Connecticut probably were covered by a nearly continuous mantle of weathered rock and saprolite, and unweathered bedrock outcrops were much less abundant than at present. That landscape would have resembled the present northern Piedmont landscape south of the glacial limit. Severely weathered bedrock (shown by map symbol x on sheet 1) occurs at many places beneath the relatively nonweathered glacial deposits and represents the remnants of the formerly widespread mantle of weathered rock. The removal of most of the weathered rock probably took place before the last (late Wisconsinan) ice sheet covered Connecticut and many differentially weathered fractures were etched in relief; however, the main features of the present landscape, including the major hills and valleys and the uplands and lowlands, were probably present in that preglacial landscape.

The character of the glacial deposits of the State is in large part determined by the physical characteristics and mineral composition of the source rocks. This is also true for alluvium and stream-terrace deposits, which are largely derived from nearby glacial deposits. The relationship holds not only in areas directly underlain by specific rocks, but also "downstream" in the direction the glacier moved. The directions of movement (shown by the arrows on figure 3) were south to southeast in most of Connecticut, but southwest to nearly west on the western side of the Connecticut Valley ice lobe. Abundant fragments of a particular rock generally occur within a few miles downstream from its source area, but scattered fragments, particularly of hard rocks, may occur tens of miles away. In the Glastonbury area of central Connecticut, till that is largely derived from Jurassic sedimentary rocks was deposited 0.3 to 1.6 km (0.2 to 1.0 mi) downstream on metamorphic rocks (Langer, 1977). Fragments of distinctive rock types downstream from restricted source areas form indicator fans, two of which are shown in figure 3. The indicator fan in western Connecticut is derived from the sedimentary and igneous rocks of the Pomperaug Valley (Pessl, 1970); the fan in southeastern Connecticut is derived from the Preston Gabbro (Goldsmith, 1982).

The following comments about the effects of specific rock types are based mostly on general impressions rather than on detailed studies. These comments apply most strongly to glacial tills; the removal of fine particles from water-washed sediment diminishes some of the contrasts in texture and color. The sedimentary rocks of central Connecticut produced glacial deposits that are commonly reddish brown; siltstones contributed considerable fine-grained material, and conglomerates contributed rounded pebbles and cobbles that have been broken out of matrix. Basalt and dolerite form narrow linear ridges of relatively hard rock within the area underlain predominantly by sedimentary rock; basalt and dolerite fragments are abundant near their ridge sources, and occur mainly in the reddish-brown deposits derived from the adjacent sedimentary rocks. Deposits in the Eastern and Western Highlands derived from quartzites contain abundant quartz sand grains and quartzite fragments. Quartzites, because of their hardness, occur as very widespread erratic fragments. Marbles produced fine-grained, light-colored, highly calcareous glacial deposits. Dark schists and phyllites produced fine-grained, dark-colored deposits. Abundant flakes of muscovite are very conspicuous in deposits derived from muscovite schists. The sulfidic schists and gneisses containing weathered iron sulfides produced very rusty deposits which locally have been cemented by iron. Granitic rocks that have few dark minerals produced light-colored deposits containing abundant quartz and feldspar in the sand and coarse silt sizes. Dark mafic and ultramafic rocks produced dark-colored deposits that contain abundant dark iron minerals. The undivided light- to medium-gray schists and gneisses underlie more than half of Connecticut; they are the source rocks for the widespread sandy-silty tills of various shades of light and medium gray and yellowish gray; associated sand and gravel are generally yellowish to light brownish gray. These widespread deposits show a considerable range of composition, color, and texture, but lack the distinctive lithologic effects produced by the other rock types mentioned above.

Glaciation

During the last (late Wisconsinan) glaciation (25-20 ka), a sector of the Laurentide ice sheet of northeastern North America spread across the St. Lawrence River valley and the Green and White Mountains of Vermont and New Hampshire, covered all of Connecticut, and reached its maximum extent on Long Island (Sirkin, 1982). Ice-movement directions are indicated by striations and grooves on bedrock, drumlin axes, and, inferentially, by the positions of ice margins during retreat (figures 3 and 4 on sheet 2). Ice movement across the State was dominantly from north-northwest to south-southeast. The principal departure from that general trend was a prominent lobation in and adjacent to the Central Lowland. On the western side of that lobe, which probably became accentuated as the ice thinned during retreat, directions of movement were to the southwest or even to the west. Weaker lobate patterns occurred in the valleys of the Quinebaug, lower Connecticut, and Housatonic Rivers, and ice movement in westernmost Connecticut was influenced by the large lobe in the Hudson River valley. The glacial meltwater deposits so conspicuous on the map were all deposited during retreat of the late Wisconsinan ice sheet.

Less is known of the earlier (probably Illinoian) glaciation recorded by the presence of a lower till. Drumlins are composed dominantly of lower till, and their directions are probably partly inherited from the earlier glaciation. Glacial meltwater deposits of this earlier glaciation are rare; they evidently were eroded or buried during the late Wisconsinan glaciation. Still earlier continental glaciations, recorded by deposits in the mid-continent and confirmed by oxygen-isotope studies in ocean sediments and Greenland ice cores (Imbrie and others, 1984; Mix, 1987; Paterson and Hammer, 1987), probably also affected Connecticut, but no direct evidence has yet been found within the State.

Glacial Ice-Laid Deposits

Till Deposits

Two glacial tills, distinctive in character and different in age, are present in Connecticut and are described in the Description of Map Units. These tills are not shown as separate units on the map because the lower, older till (also called "drumlin till") (Stone, B.D., 1989; Melvin and others, 1992) occurs almost entirely in the subsurface; generally the lower till is at the surface today only in the floors of artificial excavations that are too small to show at this map scale. The lower till may be at the surface in the upper parts of some drumlins, but its occurrence is known only from local exposures and its extent cannot be predicted; most commonly, lower till in drumlins is mantled by thin, upper till. Numerous artificial exposures and subsurface well and test-boring data indicate that lower till comprises the bulk of material within drumlins and other areas mapped as thick till (tt). Localities where lower till is (or has been) exposed are shown by an open diamond symbol on the map; such localities have now been identified in all parts of the State, in contrast to their limited known extent three decades ago (Schafer and Hartshorn, 1965; Pessl and Schafer, 1968; Pessl, 1971). The upper, younger till (also referred to as "surface till") is generally less than 4.6 m (15 ft) in thickness and comprises most of the surficial material in areas mapped as thin till (t). Localities where the two tills are (or have been) exposed in superposition are shown on the map by a solid diamond symbol. Till is shown as two separate blocks on the correlation chart to reflect its two different ages.

End Moraine Deposits

Recessional end moraines were first mapped in southeastern Connecticut by Goldsmith (1960, 1962, 1964) who defined five belts of segmented moraines. They are the Clumps, Mystic, Rocky Hollow, Ledyard, and Oxoboxo moraines, most of which extend eastward into Rhode Island (Schafer, 1961; Goldsmith, 1982). Farther west, the Old Saybrook and Madison moraines were mapped by Flint (1971, 1975); however, further field investigation revealed that the distribution of some of those moraine segments was exaggerated. We reduced their areal extent on the map and mapped separately the Hammonassett moraine in Clinton. Mapping of additional moraine segments between the two areas during the State map compilation allowed linear correlation of the Old Saybrook with the Wolf Rocks moraine (owm) in North Kingston, R.I., the Hammonassett with the Ledyard moraine (hlm), and the Madison with the Oxoboxo moraine (mom). Offshore mapping of moraine segments and of subbottom, continuous, linear ridges of proximal lacustrine fans (lcf), deposited at the grounding line of the ice margin in glacial Lake Connecticut, allowed linear correlation of the Old Saybrook moraine along the Lordship lacustrine fan ridge to the Norwalk Islands moraine in southwestern Connecticut.

The moraine belts are relatively linear, but show down-ice topographic deflection where they cross valleys. Accumulations of rock debris were concentrated in the shear zone where active ice rode up over thin stagnant ice at the margin. When the shear zone remained in one position for a significant time, concentrations of debris built up within and on top of stagnant ice (Goldsmith, 1982). This material was later deposited on the land surface by ablation processes. The linear trend of the moraine belts reflects the former position of the shear zone some relatively short distance behind the more ragged margin of stagnant ice. The segmented nature and local boulder-lag character of these moraines is probably due to the action of meltwater in the marginal zone. The moraine segments are most obvious in the upland areas between valleys. In valleys, the morainic material may be buried by meltwater deposits; in most places, meltwater deposition dominated in the valley and the morainic position is represented by the ice-proximal head of a morphosequence. Locally, moraine segments, which are more lobate than in upland areas, stand at the proximal heads of meltwater deposits in the valley.

The coastal Connecticut moraines are parallel to a much larger moraine belt that includes the Charlestown and Fishers Island moraines in Rhode Island and New York, and the Harbor Hill moraine on the north shore of Long Island (Sirkin, 1982). Parts of the Harbor Hill-Fishers Island-Charlestown moraine are shown on the Connecticut map because the area is included on the topographic base and this moraine provided basin closure for the containment of glacial Lake Connecticut in the Long Island Sound area.

A few scattered minor moraines have been identified elsewhere in the State to the north; these are much less extensive and reflect more irregular ice margins than those near the coast. One of these, the moraine at Windsorville in the Broad Brook quadrangle, is the only one in the State to have had an exposure of well-developed ice-thrust structures (Stone and others, 1982).

Glacial Meltwater Deposits

The shrinkage of the late Wisconsinan ice sheet and the retreat of its margin from south to north across Connecticut were accomplished as the ice melted faster than it was resupplied by movement from the north. The meltwater picked up rock debris carried by the ice and deposited most of it shortly beyond the ice margin. The deposits are sorted, stratified layers of gravel, sand, silt, and clay; these sediments accumulated in streams and lakes, large and small, that were fed by the meltwater. Because meltwater largely flowed in valleys during deglaciation, meltwater deposits are concentrated in those valleys and in many places are more than 30 m (100 ft) in thickness. The drift thickness map (figure 2 on sheet 2) shows that the thickest glacial deposits in Connecticut are in the deepest parts of bedrock valleys and in the Long Island Sound basin; these thick deposits are largely meltwater sediments that accumulated in glacial lakes. Figure 4 (sheet 2) shows the positions of major glacial lakes and selected ice margin retreatal positions across Connecticut.

Dominance of Glaciolacustrine Deposition

Most of the meltwater sediments in Connecticut were deposited in or graded to large and small glacial lakes. On R.F. Flint's 1930 map of the glacial geology of Connecticut, most of the meltwater deposits are mapped as "sand and gravel deposits in local temporary lakes (dammed by ice and controlled by spillways)" (Flint, 1930). Many of the different ideas about the retreat of the last ice sheet in Connecticut were mainly efforts to explain the abundance of deltaic deposits, especially in southerly sloping valleys (Gulliver, 1900; Flint, 1930, 1932, 1934; Lougee, 1938, 1953; Lougee and Vander Pyl, 1951; Black, 1977, 1982).

This report reflects our concurrence with Flint's early observation of pervasive deltaic bedding in these deposits, though not with his regional stagnation model for the formation of glacial lakes. Of the 204 map units of correlated meltwater deposits, 183 consist of deposits that were laid down in or graded to glacial lakes. These units are grouped into four types of glaciolacustrine depositional systems (major ice-dammed lakes, major sediment-dammed lakes, related series of ice-dammed ponds, and related series of sediment-dammed ponds). We include within glacial-lake units not only lake-bottom sediments and deltaic deposits, but also the fluvial deposits laid down in tributary valleys by meltwater streams that fed deltas in the lake. This leaves only 21 of the 204 units as glaciofluvial units grouped into two types of glaciofluvial depositional systems (proximal meltwater streams and distal meltwater streams).

Morphosequence Deposition and Stagnation-Zone Retreat

The glacial meltwater deposits of Connecticut resulted mainly from the interaction of three factors: (1) the form of the landscape across which the ice was retreating, (2) the form of the margin of the retreating ice, and (3) the locations of the principal meltwater streams emerging from the ice. These factors are not independent of one another—the form of the landscape influenced the other two to a considerable extent. The character of these deposits supports their interpretation through two closely related concepts: morphosequence deposition and stagnation-zone retreat (Currier, 1941; Jahns, 1941; Koteff, 1974; Koteff and Pessl, 1981). These concepts have roots more than a century old (Upham, 1878; Emerson, 1898; Woodworth, 1898), were articulated in present form more than five decades ago, and have since been exemplified in many of the quadrangle studies used in compiling this map.

Morphosequences are the basic mappable geographic-chronologic units of glacial meltwater deposits. They are the bodies of sediment formed in particular valleys during specific short periods of time as meltwater streams aggraded their beds, filled proglacial ponds and lakes, and built up to maximum levels that were commonly controlled by spillways, older deposits, or remnant ice downstream.

Seven types of morphosequences occur in Connecticut and are described in Appendix 1. They are defined by the distribution of glaciofluvial, glaciodeltaic, and glacial lake-bottom sedimentary facies, and by whether or not their heads were in contact with the ice. Near-ice-marginal morphosequences were formed short distances in front of the ice margin and were separated from it by valley segments that were traversed by meltwater streams without deposition taking place. Ice-marginal morphosequences were deposited in contact with the ice and their heads are marked by more or less well-developed ice-contact scarps (shown by hachured ice-marginal lines on the map).

The depositional heads of many ice-marginal morphosequences probably extended well up onto the edge of the ice or into tunnels within it, but melting of adjacent and subjacent ice generally destroyed such headward parts, or caused them to be collapsed downward and later buried. Most morphosequences are 0.8 to 3.2 km (0.5 to 2 mi) long, and few extend downstream more than 8 km (5 mi). Meltwater streams either terminated in deltas, or their regimes changed downstream from aggradational to balanced or degradational. Meltwater terraces probably represent the downstream parts of stream systems near the heads of which active aggradation was occurring. The ending of deposition of one morphosequence and the beginning of another probably occurred by diversion of the meltwater flow because of such events as the opening of a new and lower spillway, retreat of the margin of the ice (which was the sediment source), or the shifting of position of meltwater flow within the ice. The character of the sediment, lengths and numbers of morphosequences, and regional radiocarbon (^{14}C) dating indicate that deposition of most morphosequences probably occupied periods of time in the range of 10 to 100 years.

Collapse structure (folds and faults produced by the melting of adjacent or subjacent ice) are abundantly exposed in ice-marginal meltwater deposits. In contrast, ice-thrust structures produced by the push or drag of actively moving ice are rare and have been seen only in a few local exposures. This indicates that, for the most part, the ice-marginal deposits were laid down in contact with stagnant or dead ice, too thin to transmit forward motion. The distribution of such evidence indicates that the retreating margin of active ice was fringed by a continuous or nearly continuous zone of dead ice. The outer margin of the dead zone retreated by melting; thinning of the ice near the margin caused retreat of the shear zone which separated active from stagnant ice. This process, by which the retreating active ice is always fringed with dead ice, is called "stagnation-zone retreat." A minimum measure of the width of the fringe is given by the extent of the collapsed deposits headward of the ice contact, especially by the lengths of esker segments (Koteff and Pessl, 1981). Such data, together with inferences made from the topographic setting and texture of deposits about the proximity of continuous high-standing ice, indicate that the fringe of continuous dead ice in most places was 0.4 to 1.6 km (0.25 to 1 mi) in width. The dead ice disappeared very irregularly because of differences in such factors as ice thickness, topographic position, thickness of mantling debris, and flow of meltwater through the ice. Therefore, detached ice masses of various sizes and shapes persisted well beyond the fringe of continuous dead ice.

Dissent from use of the twin concepts of morphosequence deposition and stagnation-zone retreat was expressed by Black (1977, 1979, 1982). Using as his main example the Shetucket-Willimantic basin, he urged the alternative of basinwide regional stagnation. We disagree with many of his descriptions and his interpretations of field situations and we believe that many of his conclusions are incorrect. Although a detailed rebuttal of his arguments is beyond the scope of this report, we believe that his hypothesis of basinwide stagnation is in contradiction to the great bulk of the evidence. In fact, the Shetucket-Willimantic basin seems to us to contain one of the best portrayals of morphosequence deposition and stagnation-zone retreat in eastern Connecticut. Our depiction of that area in this report reflects that belief.

Transforming Morphosequences Into Map Units

The total number of morphosequences in Connecticut is well over a thousand, ranging from several to about 25 per quadrangle. For this reason alone, they could not be shown individually on the map. Instead, they have been combined into composite units of two to twelve morphosequences, which are of common depositional setting and were formed along the same or closely related paths of meltwater flow within a particular depositional system. These composite map units, of course, represent longer time intervals than do single morphosequences.

The 204 geographically restricted map units are categorized into six types of depositional systems which formed repeatedly in time and space during late Wisconsinan deglaciation. These depositional systems are discussed and defined in the Description of Map Units. They include almost all the meltwater deposits, regardless of the great variety of local detail. Glaciolacustrine systems are represented by four of the six types: major ice-dammed lakes (IL), major sediment-dammed lakes (SL), related series of ice-dammed ponds (IP), and related series of sediment-dammed ponds (SP). They include large, open lakes that formed in broad valleys and basins (IL and SL), and related series of small lakes and ponds that formed in the narrower upland valleys and pocket basins (IP and SP); these large and small lakes formed in northerly sloping valleys and basins where they were dammed by the ice margin and in southerly sloping valleys and basins where they were dammed by previously deposited sediments. Glaciofluvial systems include meltwater streams in close contact with (proximal to) the ice margin (FP) and meltwater streams that incised and redeposited material far from (distal to) the ice margin (FD). The six depositional systems are defined by lithostratigraphic principles; each has a typical spatial distribution of various sedimentary facies and types of morphosequences. The depositional systems are distinguished on the map by contrasting groups of colors.

Extended Ice-Margin Lines

The ice-contact scarps of ice-marginal meltwater deposits mark temporary, local positions of ice margins. The deposits generally are controlled by local topographic and hydrologic factors; therefore, they correlate from one valley to another only where lacustrine deposition in one valley required ice-marginal barriers in a nearby valley. However, ice-margin lines can be extended laterally from deposits on the basis of inferred ice-marginal trends and inferred slopes of the ice surface. General ice-marginal trends are indicated by the positions of deposits against the topography. Slopes of the ice surface are estimated to be in the vicinity of 30 to 45 m/km (100 to 150 ft/mi), substantially less than the known slopes of the margin of active ice; such slopes are plausible given the extent, altitudes, and requisite ice-margin positions of the glacial lake deposits.

The method by which ice-margin lines were extended is not without error, and the accuracy of a particular position may be explained by the following rationale: the more extensive the meltwater deposits in an area, the more reliably one may extend an ice-margin line across it; the farther one extends a line across an upland with few or no meltwater deposits, the more likely there will be error; incorrect estimates of general trends will produce errors in correlation from east to west; incorrect estimates of ice-surface slopes will produce errors in the degree of topographic inflection of ice margins. Nonetheless, the extended ice-margin lines, though probably incorrect in detail, are believed to give a reasonable picture of the forms of successive ice margins during ice retreat: the general east-northeast to west-southwest trend; the great lobe in the Central Lowland and the sublobes within it; the lesser lobes in the Quinebaug, lower Connecticut, and other valleys; and the zigzag patterns across uplands. We emphasize that these lines represent the outer margins of continuous dead-ice fringes. The inner margin of the stagnant zone, although not shown, would have much less topographic inflection.

Chronology of Ice Retreat and Major Glacial Lakes

Figure 4 (sheet 2) shows selected ice-margin retreat positions across Connecticut derived from those shown on the map. The approximate dates (given in ^{14}C years) associated with four recessional ice-margin positions are estimated from a regional array of deglaciation dates from outside of the area as well as within it (Stone and Borns, 1986; Ridge and Larsen, 1990; Stone and Ashley, 1992). Radiocarbon dates in Connecticut which are relevant to the deglacial chronology are discussed and referred to in Appendix 2 and their locations are shown on the map. Positions of major glacial lakes that dominated the deglacial history of the State are shown schematically in figure 4. Deposits of each of these lakes, as well as of the multitude of smaller glacial lakes and glaciofluvial systems shown on the map, record a detailed history of ice retreat across Connecticut; a detailed discussion of all of the lakes is beyond the scope of this report. Glacial Lakes Connecticut, Middletown, and Hitchcock were the three largest and longest-lived lakes. Deposits of these lakes are extensive and are shown on the map as multiple units. The map unit descriptions refer only to the individual deposits and not to the lake as a whole; therefore, the details of lake formation, history of deposition, and drainage are discussed in this section.

Long Island Sound Basin and Glacial Lake Connecticut

Deglaciation of the Long Island Sound basin was entirely dominated by the presence of glacial Lake Connecticut (see figure 4 on sheet 2) which was impounded in the Long Island Sound basin behind the Harbor Hill-Fishers Island-Charlestown moraine (hcm). Formation of the lake began at about 19 ka when, according to Stone and Borns (1986), the ice front began to recede from the end-moraine position. Meltwater was impounded in the expanding long, narrow basin between the moraine to the south and the retreating ice margin to the north. Initially, when lake levels were highest, the impounded water was probably coextensive with a lake in Block Island Sound, and the whole system spilled across a notch in the terminal moraine at the head of Block Channel (Lewis and Stone, 1991). As the Block Channel spillway was eroded, lake levels lowered and, eventually, the surface of the inner moraine in the relatively low area between Fishers Island and Orient Point emerged. The result was the formation of a separate body of water in the Long Island Sound basin defined by Stone and others (1985) as glacial Lake Connecticut. Glacial Lake Connecticut water levels were controlled by a spillway across the lowest point on the inner moraine at the place today called The Race. From a relative initial spillway altitude of about -10 m (-33 ft), the lake gradually lowered by erosion at The Race to a final spillway altitude of about -60 m (-197 ft); the rate of this erosion was controlled by the rate of lowering of the lake in Block Island Sound to the south.

Systematic northward retreat of the ice margin through the Long Island Sound basin is recorded by sequential ice-marginal lacustrine-fan deposits (lcf) built on the lake bottom by meltwater that issued from tunnels at the grounding line of the ice. The proximal parts of the largest of these lacustrine-fan deposits are linear, ice-margin-parallel features deposited in deeper water areas along the trends of submerged extensions of the recessional moraines of coastal Connecticut.

Deltas deposited in glacial Lake Connecticut (lcms, lcp, lcjl, lcn, lcwoo, lchm, lcew, lcenh, lcnh, lcdm, lcl, lcss, lcsnw) exist along the Connecticut shoreline near the mouths of most rivers entering Long Island Sound. Delta topset-foreset contacts are paleo-water-level indicators of the lake; they occur at altitudes of 0 to 10 m (0 to 33 ft) above present sea level in the coastal deltas and indicate that, at its highest levels (when the ice margin stood at positions near the present Connecticut shoreline), glacial Lake Connecticut occupied all of the Long Island Sound basin and extended into the present river estuaries. When their present altitudes are adjusted to the regionally established glacio-isostatic rebound slope of 0.9 m/km (4.74 ft/mi) to the N. 21° W. (Koteff and others, 1988), coastal deltas record slowly lowering lake levels during their sequential construction. Because of its east-northeast trend, the ice margin first retreated onto higher ground out of the lake basin in the southeast corner of Connecticut (east of the mouth of the Thames River). Meltwater flowed from onshore ice-margin positions down bedrock valleys and built deltas into the lake near the present coastline; these deltas record the highest lake levels, which were 0 to about 2 m (0 to about 7 ft) below present sea level, which in turn projects to The Race spillway at about -9 to -10 m (-30 to -33 ft) altitude.

By about 17.6 ka, the ice sheet had established a recessional position just offshore of the present western Connecticut coast, along the Norwalk Islands moraine and its submerged eastward extension. This time-frame is indicated by correlation of the Norwalk Islands moraine with morainal positions in New York and New Jersey (Stone and Borns, 1986). Eastward along this trend, the ice margin stood at the head of an emergent delta at Lordship that marks an interlobate angle. Lobation of the ice front southeastward from Lordship, across the deep offshore extension of the Hartford basin, is marked by an extensive lacustrine-fan deposit (Lewis and others, 1988); this fan deposit appears to correlate with a line of less extensive fans and submerged moraine segments that trend northeast and pass onshore at Old Saybrook. Deltas at the mouth of the Connecticut River (lcwoo) built in front of this ice-margin position indicate that lake levels had lowered only a few meters from the initial spillway altitude of -10 m (-33 ft) by this time.

Further ice retreat progressively deglaciated shoreline areas west of Old Saybrook. A recessional position, marked by the Hammonassett-Ledyard moraine line (hlm) that passes offshore at Hammonassett Point near Clinton, appears to correlate with a line of lacustrine fans south of New Haven that mimics the lobate shape of the Lordship lacustrine-fan deposit and lies about 5 km (3 mi) to the north of it. West of Milford, the ice front also retreated out of the lake. Deltas were built into the lake directly in front of the Hammonassett moraine (hlm) and the slightly younger Madison moraine (mom) to the northwest. At this time, west of Milford, meltwater streams flowed southward down bedrock valleys and built deltas into the lake at altitudes that indicated that the lake had lowered about 5 to 7 m (16 to 23 ft).

As deglaciation progressed, the topography of the Central Lowland produced a lobe of ice extending southward from New Haven that lingered the longest in the lake. As this ice lobe retreated out of the lake, a complex of deltas was built northeast of Milford in West Haven, New Haven, and East Haven (lcdm, lcnh, lcenh); the lake had lowered about 10 m (33 ft) by this time. These deltas are extensive both on land and offshore. In the deep central part of the lake basin, much of the varved lake-clay section, which is commonly 100 m (328 ft) in thickness, settled out in the lake concurrent with deposition of the New Haven area deltas; their stratigraphic equivalence is clearly indicated by continuous internal reflectors on seismic-reflection profiles. Onshore evidence indicates that this delta building took place during the time of ice retreat from the Lordship position to about 16 km (10 mi) north of New Haven (see 16.5 ka ice position on figure 4, sheet 2). Current estimates of deglacial chronology suggest that this period was perhaps on the order of 1,000 years.

A much thinner section of varved lake clay overlies the section that is stratigraphically equivalent to the New Haven deltas and provides evidence that the lake continued to exist as the ice margin retreated northward; however, the coarse-grained sediment supply to the lake was largely cut off when the ice margin left the shorter drainage basins feeding the lake and when new glacial lakes trapped much of the sediment in larger drainage basins to the north. During this time, the level of glacial Lake Connecticut continued to lower due to erosion at the spillway. As shoreward portions of the lakebed were subaerially exposed, streams locally entrenched older, higher level delta deposits and redeposited coarser material farther out into the lake basin; this material is seen in places at the very top of the varved lake-clay section. The gradually shrinking glacial lake may have lasted another 1,000 years; this seems a reasonable estimate for the duration of varve deposition in the upper lake section. If so, by about 15.5 ka, the lake was completely drained and the lakebed was subaerially exposed.

Eastern Highlands

The Eastern Highlands include the Quinebaug-Southeast Coastal basin, the Thames basin, and the Lower Connecticut basin (figure 1 on sheet 2). During early deglaciation of the southeast coastal area, the ice margin emerged from glacial Lake Connecticut and retreated onto land. The resultant change in the ice-flow regimen and the slow ice-margin-retreat rate in the coastal upland area resulted in deposition of three recessional moraines: the Mystic (mm), Old Saybrook-Wolf Rocks (owm), and Hammonassett-Ledyard (hlm) moraines. Concurrent with early morainic deposition, fluviodeltaic deposits were laid down and graded to glacial Lake Connecticut in south-draining coastal valleys. These early deposits served as initial sediment dams for several series of ponds (SP depositional system) that developed with continued northward ice retreat. Once the ice margin had retreated to north of the east-west-trending Quinebaug River drainage divide, several series of small ice-dammed ponds (IP depositional system) as well as the larger ice-dammed glacial Lakes Voluntown, Pachaug, and Oneco (IL units lvo, lp, lon) were dammed by the edge of the ice. With further retreat into the broad Quinebaug River valley, sediment-dammed lake deposition (SL) began with the formation of glacial Lake Quinebaug (lqb) (first described by Stone and Randall, 1978) which controlled meltwater deposition for a considerable time during ice retreat in the valley northward to the vicinity of Danielson. This lake was impounded by a glacial drift dam that filled the bedrock gorge of the present Quinebaug River south of Jewett City. The water level of the lake was controlled by a spillway over bedrock adjacent to the gorge with an altitude of 41 m (134 ft). The glacial lake lengthened northward in the Quinebaug lowland as the glacier margin retreated. Deltaic deposition controlled by the glacial Lake Quinebaug water plane was interrupted for a time by the deposition of Quinebaug River valley deposits (qb) as small, sediment-dammed ponds (SP) temporarily controlled deltaic deposition at levels slightly higher than glacial Lake Quinebaug; however, after this brief interval, erosion of sediment dams farther south returned the controlling base level to the glacial Lake Quinebaug water plane. During ice retreat in the Quinebaug valley northward from Danielson, SP depositional systems controlled the levels of predominately ice-marginal deltaic and fluviodeltaic deposits (including units da, pt, ew, wt, and fr). In small, north-draining tributary valleys, series of small, ice-dammed ponds (IP) controlled deltaic deposition for five units (mb, wr, wb, epg, and nth).

In the Thames basin, early meltwater deposits in south-draining coastal streams were graded to glacial Lake Connecticut. In the lower Thames River valley, these deposits were largely removed by later meltwater erosion. Moraine segments also locally cross the valley. Glacial Lake Uncasville was impounded as a sediment-dammed lake, probably behind a segment of the Ledyard moraine (later eroded) in the Thames River valley. Ice-marginal deltas and fluviodeltas built into this lake during the time of ice retreat northward to the Norwich area. Farther north in the basin, SP depositional processes prevailed in the relatively narrow Willimantic, Yantic, and Shetucket River valleys where seventeen map units have been differentiated (mr, bg, sk, hp, mf, na, wru, wrl, wil, lru, lrl, mbb, str, yr, tc, ox, el). The most laterally extensive SP deposits (wil) were laid down in a series of small, sediment-dammed ponds associated with 12 successive ice-marginal positions in the Shetucket River valley. In the small northerly draining tributary valleys to the larger southerly draining valleys, IP depositional processes controlled deposition of six map units (mp, cr, cs, fv, gl, led).

South-draining valleys in the Lower Connecticut basin were initially blocked by meltwater deposits graded to glacial Lake Connecticut and locally by moraine segments. This process was succeeded by sequential sediment-dammed ponding (SP) behind the coastal-area deposits; 12 units of sediment-dammed pond deposits were differentiated in the Lower Connecticut basin. Of these units, lcc, lce, and lct in the Connecticut River valley were the most extensive, completely filling the valley from side to side. The deposits formed the lengthy sediment dam for glacial Lake Middletown. Major ice-dammed lakes (IL), glacial Lakes Essex (lex) and Colchester (lc), formed in two north-draining valleys in the Lower Connecticut basin. Five series of small ice-dammed ponds (IP) controlled deposition in smaller north-draining valleys (whd, mop, jg, cl, bv).

Because of the overall east-northeast trend of the retreating ice margin, the Eastern Highlands were entirely deglaciated while ice still remained in the Western Highlands and Central Lowland. Ice retreat across this area probably occurred between 18.0 and 16.0 ka based on regional correlation of ice margins (Stone and Borns, 1986). The earliest ^{14}C-dated organic horizons in the area, 15.0 ka at Rogers Lake (Davis and others, 1980) and 15.2 ka at Cedar Swamp (McWeeney, 1995), both record tundra vegetation at that time, but at both sites cores did not penetrate the base of the organic section.

Western Highlands

The Western Highlands include the Housatonic-Southwestern Coastal basin and the Naugatuck basin (figure 1 on sheet 2), and are dominated by narrow bedrock valleys within extensive upland areas. Because of the narrowness of most valley segments in this area, meltwater deposits are predominantly units of the SP and IP depositional systems. Several valleys contain proximal fluvial deposits (FP) in their steeper reaches. Ten major glacial lakes developed in the broader valley segments as the ice margin retreated northwestward through the area. Once the ice margin had retreated across the east-west-trending divide that separates the southwestern coastal rivers from the Housatonic drainage basin, glacial Lakes Danbury and Pootatuck formed as ice-dammed lakes (IL) in north-draining tributary valleys following initial deposition in a series of ice-dammed ponds (IP system). Glacial Lake Danbury, first described by Thompson (1975), was the most extensive lake in the Western Highlands; it existed in the Still River valley and is recorded by deposits (lds, ldp, and ldh) graded to three successively lower stages and by extensive lake-bottom deposits. Glacial Lake Pomperaug (lpg) formed as a sediment-dammed lake (SL) in the Pomperaug valley, a south-draining tributary valley to the Housatonic River; it was dammed by sediment in the Housatonic valley and coexisted with the Pumpkin Hill stage of glacial Lake Danbury. North-draining tributary valleys in the Western Highlands contain deposits of ten series of ice-dammed ponds (IP depositional system). Ice retreat in the relatively narrow Housatonic and Naugatuck River valleys was dominated by sedimentation in a series of small sediment-dammed ponds (SP depositional system). Locally, in the more steeply sloping valley reaches, proximal glaciofluvial deposition (FP) took place. Continued flow of distal meltwater, especially in the Housatonic River valley, eroded large parts of the ice-marginal deposits, leaving only remnant terraces on valley sides.

With retreat into the northwest corner of Connecticut, the ice margin dammed glacial Lake Hollenbeck in the north-draining Hollenbeck River valley and glacial Lake Norfolk in the Blackberry River valley. Deposition in these north-draining valleys largely preceded development of glacial Lake Great Falls. The lake developed in the broad, overdeepened basin of the Housatonic River valley, which is underlain by easily erodible marble. Ice-marginal deltas built from the last upland ice-margin positions in Connecticut were constructed in glacial Lake Great Falls; deposits of this lake are extensive in Massachusetts.

Central Lowland

The Central Lowland includes the Farmington-Quinnipiac basin and the Upper Connecticut basin (figure 1 on sheet 2). Deglaciation of this area was characterized by major lobation of the ice margin which was strongly influenced by the topographic relief between the Eastern and Western Highlands and the Central Lowland. The trap-rock ridges that separate the Farmington-Quinnipiac basin from the Upper Connecticut basin also produced a topographic inflection in the lobate ice margin. Ice retreat was to the northwest on the east side of the lowland and to the northeast on the west side (see figures 3 and 4 on sheet 2). Meltwater deposition in the Central Lowland was dominantly controlled by major glacial lakes. In the Farmington-Quinnipiac basin, initial damming of the Quinnipiac valley was provided by part of the New Haven deltaic deposits (lcnh) graded to glacial Lake Connecticut; these deposits filled the valley at Fair Haven and impounded glacial Lake Quinnipiac (lq). This lake received only local ice-marginal deltaic deposition, but extensive lake-bottom deposits, New Haven Clay (Flint, 1933), filled the Quinnipiac River valley to just south of the bedrock gorge at South Meriden. North of the gorge, glacial Lakes Southington (ls), Farmington (lf), and Tariffville (lt), all major sediment-dammed lakes (SL depositional system), controlled meltwater deposition in the basin. In narrower, steeper parts of the valley, SP depositional processes prevailed during northward ice retreat and units sb, un, ws, sl, and ss were deposited in the valley. In north-draining tributary valleys to the main valley, the ice margin dammed major glacial Lakes Bristol and Nepaug (IL) and six series of small, ice-dammed ponds (IP).

Following drainage of glacial lakes in the Farmington-Quinnipiac valley, an extensive glaciofluvial unit, the Quinnipiac River valley terrace deposits (qt) of the FD depositional system, was deposited and inset into the older lake deposits. This unit is the most extensive glaciofluvial deposit in the State (Stone and others, 1985) and was deposited by distal meltwater that flowed from the Western Highlands southward down the Quinnipiac River valley. The meltwater carried light-colored, mica-rich, quartzofeldspathic sands derived from crystalline rocks. The sands contrast in color with the red-brown color of the ice-marginal, glacial-lake sediments derived from local Mesozoic rocks in the valley (Krynine, 1937; Lougee, 1938). When the Tariffville gap, where the present Farmington River flows through the trap-rock ridge into the Upper Connecticut basin, was deglaciated, meltwater drainage was diverted northward and deposition of the terraces ceased.

Deglaciation of the upper Connecticut valley was dominated by sedimentation in glacial Lakes Hitchcock and Middletown. Several ice-dammed lakes in north-draining tributary valleys of the Connecticut River valley preceded these two lakes. Glacial Lake Coginchaug was impounded in the north-draining Coginchaug River valley. The early stage of the lake (lcgd) spilled across the main drainage divide to the south; the later stage of the lake (lcgm) spilled eastward into the Connecticut valley. Deposition of the Hanging Hills unit (hh of the IP depositional system) also preceded formation of glacial Lake Middletown in Meriden in small northeast-sloping valleys which drain the dip slope of the Hanging Hills (trap-rock ridges) and are tributary to the Mattabesset River.

On the east side of the upper Connecticut valley, north-draining valleys contained a series of ice-dammed ponds (cd) which spilled across the upper Connecticut drainage divide. They were followed by a series of long, narrow, ice-dammed glacial lakes that formed in valleys oblique to the trend of the ice margin: glacial Lakes Roaring Brook (lrb), Salmon Brook (lsb), and Manchester (lma). These ice-dammed depositional systems formed in north-sloping tributary valleys to the Upper Connecticut basin and preceded the development of major sediment-dammed lakes in the main basin.

Glacial Lake Middletown (SL)

Glacial Lake Middletown first developed along the Connecticut River and in the Mattabesset River basin. The lake was impounded by a long mass of earlier deposits (lcc, lce, lct) in the lower Connecticut River valley at and south of The Straits; the spillway, with an initial altitude of about 40 m (130 ft) was over these deposits. Successive ice-marginal deltaic deposits were built into the lake as the ice retreated northward. When adjusted for the regionally established postglacial tilt of 0.9 m/km (4.74 ft/mi) to the N. 21° W., delta topset-foreset contacts indicate that the lake slowly lowered due to erosion of its sediment dam. Glacial Lake Middletown occupied the Middletown basin in the lower Mattabesset valley and extended into the Berlin basin in the upper Mattabesset valley, as indicated by accordant delta levels, by basin geometry resulting in ice-margin positions that trend northwest-southeast, and by the extent of clays in the Berlin area. Deltas in Cromwell (lmc), Newington (lmn), and New Britain (lmw) were built contemporaneously and record lake levels at the spillway of about 34 to 35 m (110 to 115 ft).

Just north of the Cromwell deltas, deltas of the Dividend Brook deposits (db) were laid down in waters that were temporarily ponded to a higher level than glacial Lake Middletown and were controlled by the Dividend Brook spillway over Cromwell deltaic deposits (lmc); this spillway was not eroded lower than its present level of 39 m (129 ft) because of the presence of glacial Lake Middletown at its mouth.

When the ice uncovered the lower part of the divide between the Hartford basin and the Middletown-Berlin-New Britain basin, where the New Britain spillway of glacial Lake Hitchcock would later exist, glacial Lake Middletown persisted at a level high enough to spread across the divide into the Hartford basin. When the ice retreated from the north end of Cedar Mountain (Newington-Hartford town line), the Dividend Brook spillway was abandoned and glacial Lake Middletown spread eastward into the southern end of the basin later occupied by glacial Lake Hitchcock. Deltaic deposits (lmw, lme, lmg, and lmwv) as well as lake-bottom deposits (lmb) in the Hartford basin all occur at altitudes accordant with glacial Lake Middletown, but too high to have been controlled by any possible early level of the New Britain spillway. Not until glacial Lake Middletown had lowered to below 34 to 35 m (110 to 115 ft) at the divide (about 20 m (65 ft) at The Straits spillway) could the New Britain spillway come into use as the outlet for glacial Lake Hitchcock.

Glacial Lake Hitchcock (SL)

Glacial Lake Hitchcock existed in the upper Connecticut River basin in Connecticut, Massachusetts, Vermont, and New Hampshire, lengthening to at least 298 km (185 mi) as the ice retreated northward to the vicinity of Burke, Vt. The Connecticut River valley was dammed to an altitude of 46 to 49 km (150 to 160 ft) in the vicinity of Rocky Hill and Glastonbury by deposits of glacial Lake Middletown (lmc and db); this mass of stratified drift is often referred to as “the Rocky Hill dam.” The spillway for glacial Lake Hitchcock was not over the dam, however, but at the lowest place across the Mattabesset River drainage divide between the Hartford basin and the Middletown-Berlin basin in New Britain. When the ice margin first retreated into the Hartford basin, north of that divide, glacial Lake Middletown water covered the later New Britain spillway location and early ice-marginal deltas in the Hartford basin were controlled by glacial Lake Middletown. Not until glacial Lake Middletown had dropped to below 35 m (115 ft) could the New Britain spillway area emerge and glacial Lake Hitchcock exist as a separate water body; this occurred at about the time that the ice margin was at Windsor and East Windsor.

During the early life of glacial Lake Hitchcock, the New Britain spillway was eroded into till and older stratified drift so that water levels at the spillway dropped from about 35 m (115 ft) down to 25 m (82 ft) in altitude (Langer, 1977; Langer and London, 1979). In Connecticut, all ice-marginal and distal-meltwater-fed deltas, as well as one small delta built by meteoric water, record lake levels higher than the longer lived stable level. These deltas show a gradual lowering of the lake level as the ice retreated northward and the New Britain spillway was incised down to bedrock. Ice-marginal deltas in Windsor (lhhw) and East Windsor (lhhe) record 34- to 35-m (110- to 115-ft) levels at the spillway. To the north, ice-marginal deltas in Suffield (lhhr) and Enfield (lhhs) indicate 32- to 33-m (105- to 110-ft) levels at the spillway; still farther north in Suffield and Enfield, the Shea Corner (lhhsc) and Enfield (lhhrn) deltaic deposits record levels just below 30 m (100 ft) at the New Britain spillway. This early phase of glacial Lake Hitchcock is recorded by ice-marginal deltas that are found well into southern Massachusetts and that were built to lake levels between 26 and 29 m (85 and 95 ft) at the spillway. This higher-than-stable-level phase of the lake is referred to as the "Connecticut Phase" (Koteff and others, 1988). It is important to note that deepening of the spillway channel was controlled by conditions 48 to 64 km (30 to 40 mi) to the south (the New Britain spillway was an independent control for lake levels). The base level for waters exiting the spillway was controlled by downcutting in the lower Connecticut River valley and by lowering levels of glacial Lake Connecticut in the Long Island Sound basin. The Rocky Hill dam area was glacio-isostatically depressed about 44 m (145 ft) and the New Britain spillway area was depressed about 50 m (165 ft) (more than the area at the mouth of the Connecticut River). In order for the New Britain spillway to lower by 10 m (33 ft) during the early phase of the lake, glacial Lake Connecticut had to have already lowered to below -25 m (-82 ft) in altitude.

Delta levels in Massachusetts indicate that a stable lake level, 25 m (82 ft) in altitude, had been reached by the time the ice margin had retreated to just north of the Chicopee River valley; regional correlation of ^{14}C dates (Stone and Borns, 1986) place the ice front in this position at about 15 ka. The 25-m (82-ft) level indicates that the water flowing through the spillway was about 7 m (24 ft) deep because its bedrock floor today is at about 18 m (58 ft) in altitude. Altitudes of topset-foreset contacts of ice-marginal deltas, from southern Massachusetts to the lake's northernmost extent, project to the stable level (25 m (82 ft) 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 (4.74 ft/mi). The linearity of these projected delta altitudes indicates that the lake level was stable during the time of ice retreat from Chicopee, Mass., to Lyme, N.H., and that postglacial rebound of the land surface did not begin until after all ice-marginal deltas had been built, probably between 14 and 13.5 ka (Koteff and Larsen, 1989). Deltas that were not associated with the ice margin, but rather were built by meteoric water in most river valleys that entered the lake, also project to the stable lake level. In Connecticut, these include unit lhsh associated with the Hockanum River, unit lhss associated with the Scantic River, and unit lhsb, where the Farmington River constructed a large delta northeastward into the lake in the area now surrounding Bradley International Airport. The Bradley International Airport delta covers about 52 km^2 (20 mi^2) and its entire surface (which is tilted up to the N. 21° W. in the amount of 0.9 m/km (4.74 ft/mi)) is graded to the stable 25-m (82-ft) level; these two facts provide evidence for the long duration of the stable level and also indicate that the lake was not affected by glacio-isostatic tilting until after nearly all of its deltas had been constructed.

It is important also to note that the New Britain spillway could not have lowered further than the 25-m (82-ft) level. This is because the 25-m (82-ft) altitude at the New Britain spillway is equivalent to a -25-m (-82-ft) altitude at the mouth of the Connecticut River when the 50 m (164 ft) of differential depression between the two localities is taken into account. The base of the channel, through which the paleo-Connecticut River carried water that spilled from glacial Lake Hitchcock, was imposed on bedrock at -27 m (-89 ft) in altitude at the mouth of the present Connecticut River east of Saybrook Point; this point was the actual control for the "Stable Phase" (Koteff and others, 1988) of glacial Lake Hitchcock. The "Stable Phase" of glacial Lake Hitchcock lasted from about 15 ka until about 13.7 ka; during this time, the southern part of the basin (south of the Holyoke Range in Massachusetts) was largely filled with deltaic and lake-bottom sediments. Preserved lake-bottom surfaces in Connecticut are at about 14 m (45 ft) in altitude in the south and 44 m (145 ft) in the north; the tilted stable-level paleo-waterplane over this area is at 19 m (63 ft) in altitude at the north edge of the Rocky Hill dam and 52 m (172 ft) at the Massachusetts border; thus, toward the end of the "Stable Phase" before the dam was breached, water depths in the lake were only 6 to 8 m (20 to 25 ft). Because the bedrock basin that contained the lake north of the Holyoke Range in Massachusetts is deeper, the lake was not filled with sediment to the extent that it was in the southern basin. North of the Holyoke Range in Massachusetts, preserved lake-bottom surfaces are at 46 m (150 ft) in altitude, and at the end of the "Stable Phase," water depth was about 46 m (150 ft).

Fluviodeltaic deposits (ft and lhf) built southeastward into the lake by the Farmington River record a "Post-stable Phase" (Koteff and others, 1988) of the lake during which levels were lower than the 25-m (82-ft) level at the New Britain spillway. A topset-foreset contact in the lhf deltaic deposits north of the Farmington River is at 39 m (127 ft); delta-surface altitudes in the same unit to the south of the river indicate slightly lower water levels. These levels project southward below the New Britain spillway level to 15 to 18 m (50 to 60 ft) in altitude at the Rocky Hill dam and record lowering of lake levels as the dam was entrenched. A preserved 17-m (55-ft) terrace inset into the Rocky Hill dam sediments on both sides of the present Connecticut River in Rocky Hill and Glastonbury records this "Post-Stable Phase" which was relatively brief in Connecticut. A ^{14}C date of 13,540±90 B.P. (Beta-59094, CAMS-4875) on plant debris in lacustrine sands at the top of the lake-bottom section (radiocarbon-dated locality 10, Appendix 2) associated with the Farmington River deltaic deposits (lhf) establishes that the time of dam breech was at about 13.5 ka.

The dam most likely was 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, much less drain, until its bed was raised by glacio-isostatic tilting. Dam breaching and initiation of isostatic rebound was required in order to establish the lower water-level altitudes recorded in the "Post-Stable Phase" Farmington River deltaic deposits (lhf). Once this process began, it proceeded rapidly as the dam was incised from just above 18 m (60 ft) in altitude (the stable level at the dam) to just above 12 m (40 ft); once this 6 m (20 ft) 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 (along the terraces of unit st) over the 80-km (50-mi) 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 (130 ft) (lowered from stable level by only 6 m (20 ft)); continued lowering of the lake was controlled by the rate of rebound, which made it possible for the lake bed south of the Holyoke Range to be incised.

An approximate 4,000-year life span for glacial Lake Hitchcock was indicated by Antevs (1922) through a method of correlating varves in clay pits from Hartford, Conn., to the north end of the lake basin in St. Johnsbury, Vt. 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 to construct Antevs' (1922) New England varve chronology between varve-year 3,001 and varve-year 7,000. Recently, Ridge and Larsen (1990) fit a 533-year varve section from Canoe Brook in southern Vermont into the relative varve chronology of Antevs (1922); they also placed the chronology in an absolute time frame with a 12.4 ka ^{14}C date on plant debris in the Canoe Brook section at the position of varve 463 (varve 6,150 in the Antevs chronology). Using this calibration of the varve chronology, lacustrine deposition at the south end of glacial Lake Hitchcock (varve 3,001) began at about 15.5 ka. The early Connecticut phase was followed by the longer "Stable Phase" of the lake which lasted until about 13.5 ka (varve 5,050). The "Post-Stable Phase" of the lake, which lasted only briefly in Connecticut, continued for another 2,000 years north of the Holyoke Range until about 11.5 ka (varve 7,000) (Stone and Ashley, 1992, 1995; Stone, 1999).

Postglacial Conditions

Postglacial deposits in Connecticut include stream-terrace (st), talus (ta), dune (d), flood-plain alluvium (a), swamp (sw), salt-marsh (sm), beach (b), fluvial-estuarine channel-fill (ch), and marine delta (md, mdd) deposits; the onset of postglacial conditions was time-transgressive and began several thousand years earlier in the southern part of the State than in the northern parts.

In the Long Island Sound basin, significant postglacial events include the drainage of glacial Lake Connecticut and subsequent sea-level rise. The remnant glacial lake was probably completely drained by 15.5 ka and a fluvial channel system (linear scarp symbol on map) was being carved on the lake floor by meteoric streams flowing to the south in coastal Connecticut, and to the north on the north shore of Long Island; these tributary channels joined a major east-west-trending trunk channel which also received distal meltwater drainage from the Hudson River valley to the west (Stanford and Harper, 1991). The channel system exited the basin through the lake-spillway notch in the end moraine at The Race and provided a path through the moraine for the subsequent transgression of the sea from the south (Lewis and Stone, 1991). Minor fluvial sediments were deposited in the bottoms of the channels during the time that they were occupied by streams. The channels are filled predominantly with estuarine sediment (ch) deposited as the early postglacial sea flooded these low-lying areas of the drained lake basin when eustatic sea level began to rise significantly between 16 and 15 ka (Fairbanks, 1989; Bard and others, 1990) and before glacio-isostatic rebound began.

A major wave-cut marine unconformity (labeled on seismic section *C–C′*) was cut across the top of the estuarine channel fill and over higher lake deposits as sea level rose. The marine unconformity is present in seismic sections up to altitudes of about -25 m (-82 ft), indicating that sea level probably rose to this height in central Long Island Sound before crustal rebound began.

Figure 5 (sheet 2) shows a conceptual relative sea-level curve for central Long Island Sound (highlighted line). The curve was derived by combining the eustatic sea-level curve from Barbados (Fairbanks, 1989; Bard and others, 1990) with a curve representing the timing and total depth of glacio-isostatic depression in central Long Island Sound. The uplift curve is based on several assumptions (see caption for figure 5) that are indicated from regional evidence, some of which is presented in this report and in Koteff and Larsen (1989) and Stone and Ashley (1995). The presence of the extensive marine delta (md) that records a -40-m (-131-ft) to -50 m (-164 ft) relative sea level in central Long Island Sound (Lewis and Stone, 1991; Stone and Lewis, 1991) provides good evidence for the conceptualized relative sea-level curve. The large volume of deltaic sediment required a significant length of time for construction. The topset-foreset contact of the marine delta indicates that sea level was relatively (at -40 to -45 m (-131 to -147 ft)) stable during the deposition of the delta.

The only possible source of the great volume of sediment contained within the marine delta was the drained lakebed of glacial Lake Hitchcock in the Connecticut Valley to the north. This sediment supply became available only when the "Stable Phase" of glacial Lake Hitchcock ended at about 13.5 ka; as previously discussed, glacio-isostatic uplift had to occur in order for glacial Lake Hitchcock to drain. Regional evidence from northern New England (Barnhardt and others, 1995; Koteff, Robinson, and others, 1993; Koteff, Thompson, and others, 1995) also indicates that isostatic rebound began around this time. Thus, the early rapid rate of uplift was balanced with the equally rapid rate of eustatic sea-level rise resulting in a sea-level stand in Long Island Sound at about -40 m (-131 ft) for several thousand years (between 13.0 and 9.5 ka). During that time, the marine delta was built and the Connecticut River terrace and flood-plain surfaces were incised. Recently obtained ^{14}C dates (9,370±100 Beta-52257, 8,530±80 Beta-52256) on basal organic material beneath the lowest terrace surfaces along the Connecticut River in Massachusetts indicate that most of the postlake incision into the lakebed had been accomplished by ~9.0 ka (Stone and Ashley, 1992). The volume of eroded lakebed sediment, as calculated from the area and depth of incised terraces, is 12 billion m^3 (16 billion yd^3); this material now composes the marine delta, the calculated volume of which is 11.5 billion m^3 (15 billion yd^3).

As eustatic rise overtook the rate of isostatic rebound, relative sea level in central Long Island Sound rose continuously; the transgression submerged the marine delta and a blanket of marine mud (shown only in sections *B–B′* and *C–C′*) accumulated over the entire basin. As marine waters deepened, intense tidal-scour conditions developed in eastern Long Island Sound, resulting in the local reworking of marine-delta sediments and the development of a very large sand-wave field in that part of Long Island Sound (Fenster and others, 1990). A record of 4 to 5 m (13 to 16 ft) of sea-level rise during the last 4,000 to 5,000 years is preserved in coastal salt-marsh deposits (Bloom and Stuiver, 1963; van de Plassche and others, 1989; Patton and Horne, 1991; van de Plassche, 1991).

In most of mainland Connecticut, postglacial activity consisted predominantly of incision of glacial deposits by meteoric streams along stream-terrace surfaces, followed by the establishment of flood plains at modern levels. Streams had eroded to modern flood-plain levels relatively early, in some cases before 12.0 ka (O'Leary, 1975; Stone and Randall, 1978). Postglacial winds were intense and widespread as indicated by the ubiquitous blanket of eolian sand and silt that overlies glacial sediments throughout the State and in which the modern soil is developed. The postglacial climate was severely cold for several thousand years following deglaciation. Paleobotanical studies reveal that treeless, tundra vegetation dominated by dwarf willow (*Salix herbacia*), sedges (*Cyprus*, *Carix*), and herbs and shrubs (*Dryas*, *Artemesia*), dated from earlier than 15 ka to about 13 ka, was present in the area (Davis and others, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987; Thorson and Webb, 1991). Also, wedge-shaped features with a polygonal ground pattern, interpreted as ice-wedge casts, deform eolian-sand-capped glacial sediments in numerous localities in Connecticut (Schafer and Hartshorn, 1965; Schafer, 1968; O'Leary, 1975; Stone and Ashley, 1992). These features indicate that permafrost existed locally in areas where substrate conditions were favorable to its formation. The presence of permafrost structures indicates that mean annual temperatures were below 0°C during the early postglacial time interval.

In the upper Connecticut basin, postglacial conditions were dominated by the continued existence of glacial Lake Hitchcock several thousand years after the ice margin retreated from the area. Extensive fields of eolian sand dunes formed in the treeless environment, indicating the continued effects of strong winds. Dunes are present on the relict deltaic and lake-bottom surfaces of glacial Lake Hitchcock. Dunes on deltaic and high-level lake-bottom surfaces were formed by north to north-northeasterly paleowinds; these surfaces were available as early as 15.5 ka. Dunes on stable-level lake-bottom surfaces were formed by northwesterly paleowinds; these surfaces became available at about 13.5 ka as glacial Lake Hitchcock drained. Evidence that severely cold temperatures persisted until the time of glacial Lake Hitchcock drainage exists due to the presence of hundreds of circular to subcircular, rimmed depressions (interpreted as pingo scars) developed in the drained lakebed sediments (Stone and Ashley, 1989; Stone and others, 1991; Stone and Ashley, 1992). Paleobotanical records indicate a warming of the postglacial climate at about 12.5 ka, accompanied by reforestation of the landscape by successive spruce, pine, and hardwood forests from 12.5 to 9 ka (Davis, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987).

Description of Map Units

Map units include surficial materials more than 1 m (3 ft) thick that overlie bedrock. A discontinuous veneer of eolian fine sand and thin colluvium on slopes is not mapped. Because all geologic units are of Quaternary age, the letter "Q" is not included as a prefix to map-unit symbols.

For many units, the U.S. Geological Survey 7.5-minute 1:24,000-scale topographic quadrangle names are shown in parentheses at the end of most descriptions. See Appendix 3 for information on the sources of geologic data by quadrangle name. Some geographic features mentioned in these descriptions are shown on the source 1:24,000-scale quadrangles, but not on the 1:100,000-scale base map for this report; the source maps should be consulted in order to locate these features. In addition, some local names have been added to this base map.

The contour intervals for the 1:100,000-scale base map are in metric units; however, all measurements for onshore units were made in English units based on the 1:24,000-scale quadrangle maps, where contour intervals are in feet. Therefore, dual measurements are provided for altitudes, distances, and thickness measurements in unit descriptions. Altitudes and depths mentioned in offshore units are given in meters below mean sea level (MSL) for convenience in referencing the marine seismic-reflection profile data, which is recorded and described in metric units. The seismic-reflection profiles are the primary source of offshore geologic information.

Map units of glacial meltwater deposits are organized under the six depositional systems in which they occur, from west to east geographically by the geologic basin in which they occur. The "divides" mentioned in many unit descriptions refer to those separating these geologic basins (see figure 1 on sheet 2 for locations of the eight geologic basins).

Most unit descriptions for glacial meltwater deposits make reference to types of morphosequences that occur in that particular map unit. It is highly recommended that the descriptions of sedimentary facies and morphosequences presented in Appendix 1 be consulted before attempting to understand individual map-unit descriptions.

Postglacial Deposits (Holocene)

af **Artificial fill**—Earth and manmade materials including rocks, gravel, sand, silt, clay, concrete, and select refuse artificially and extensively emplaced, principally in coastal areas. Highway and railroad fill, areas of landfills, and local fill in urban areas are not mapped

a **Flood-plain alluvium**—Sand, gravel, silt, minor clay, and some organic material in flood plains of modern streams. Along smaller streams, texture of alluvium is commonly variable both laterally and vertically, but overall texture is often similar to adjacent glacial materials. Thickness commonly less than 2 m (6 ft). Along larger rivers, contains gravel and sand at base, overlain by laminated sand, silt, and minor clay, as much as 8 m (25 ft) thick. Alluvium of Connecticut River north of Rocky Hill is chiefly very fine sand and silt, less than 5 m (18 ft) thick. Along Connecticut River south of Rocky Hill, alluvium is chiefly fine to medium sand, as much as 12 m (40 ft) thick. Alluvium overlies glacial stratified sand and gravel, coarse gravel, or till in upland valleys; in lowlands, commonly overlies sand or silty-clayey lake-bottom deposits

sw **Swamp deposits**—Muck and peat that contain minor amounts of sand, silt, and clay overlying laminated organic silt, clay, and sand. Organic peat and muck is decomposed, fibrous and granular, woody herbaceous material. Thickness of organic materials is commonly less than 3 m (10 ft). Some deposits accumulated in poorly drained areas, mostly in shallow, low-lying basins in till and (or) bedrock areas; other deposits accumulated in relatively deep, closed depressions (kettles) in ice-proximal, glacial meltwater deposits, and in shallower depressions and swales on glacial lake-bottom surfaces; some deposits occupy low swales between alluvial levees on Holocene flood-plain surfaces. Thicker peat and muck deposits (for example, 4 m (13 ft) at Totoket bog, a kettle-hole swamp just north of Route 80 and west of Bare Plain Cemetery in the village of Totoket in the Branford quadrangle; 10 to 14 m (33 to 46 ft) beneath Linsley Pond and Cedar Pond in North Branford; 11 m (36 ft) beneath Rogers Lake in Lyme; and 4 m (13 ft) at Durham Meadows) preserve a record of vegetation changes and hence a paleoclimatic record for postglacial time. Oldest postglacial radiocarbon dates in Connecticut (~15 ka) come from base of organic swamp deposits at Totoket bog and Rogers Lake. Generally overlie materials of adjacent map unit. Shown only where greater than 25 acres in area

sm **Tidal-marsh deposits**—Peat and muck, generally 1 to 9 m (3 to 30 ft) thick, interbedded at depth with laminated fine sand and silt. Organic peat and muck is decomposed, fibrous and matted, herbaceous and silty-herbaceous material that accumulated in marshes at and upstream from mouths of streams open to marine waters of Long Island Sound; marshes include coastal salt marshes and brackish to freshwater tidal marshes farther up estuaries. Vegetation growing today in tidal marshes as well as plants preserved in peat deposits ranges from salt-water species (such as *Spartina alterniflora* and *S. patens*) to species tolerant to brackish water (such as *Typha angustiflora* and *Phragmites australis*) to freshwater sedge and rush species (such as *Scirpus fluviatilus*, *Juncas accumintus*, and *Pontederia cordata*). Vertical accretion and upstream (northward) transgression through time records drowning of former alluvial terraces along coastal streams during postglacial sea-level rise. Basal portion of most of the tidal-marsh deposits shown on map lies on surfaces above -5 m (-15 ft) mean sea level (MSL); a sea-level-rise curve generated from radiocarbon dates from tidal-marsh peat deposits along lower Connecticut River (Patton and Horne, 1991) indicates marine transgression beginning 4,000 years ago. In some places, such as marsh along Hammock River in Clinton, salt-marsh peat overlies estuarine deposits; these peat deposits are composed of organic sand and silt and freshwater sedge peat at depths of up to -12 m (-40 ft) MSL, which yield older radiocarbon dates that indicate a marine transgression beginning 7,000 years ago (Bloom and Stuiver, 1963; van de Plassche and others, 1989). Shown only where greater than 25 acres in area

ta **Talus**—Angular, loose blocks of basalt and diabase accumulated by rockfall and creep at base of bedrock cliffs along linear traprock ridges in Central Lowland. Forms steep unstable slopes. Generally less than 6 m (20 ft) thick

b **Coastal beach and dune deposits**—Fine to coarse sand and local pebble-cobble gravel in modern beach deposits. Texture of beach deposits varies over short distances and is generally controlled by texture of nearby glacial materials exposed to wave action. Beach deposits are poorly to well sorted and generally less than 2 m (6 ft) thick. Locally includes dune deposits consisting of relatively well sorted, fine to coarse sand in transverse coastal eolian dunes that are 1 to 3 m (3 to 10 ft) thick

Early Postglacial Deposits (Early Holocene to Late Pleistocene, Wisconsinan)

st **Stream-terrace deposits**—Sand, gravel, and silt deposited by meteoric water on terraces that were cut into glacial meltwater sediments. Texture is variable vertically and laterally, but is chiefly coarse pebbly sand, commonly similar to that of adjacent glacial deposits. Thickness ranges from 1 to 5 m (3 to 15 ft). Distinguished from meltwater-terrace deposition by its lower position in valley, commonly only 3 to 6 m (10 to 20 ft) above altitude of modern flood plains

d **Inland dune deposits**—Medium, relatively well sorted sand, in transverse, parabolic, and hummocky dunes as much as 12 m (40 ft) thick. Most common in drained basin of glacial Lake Hitchcock where sand was derived from extensive glacial-lake deltaic deposits. Dunes overlying lake-bottom deposits of glacial Lake Middletown and early phase of glacial Lake Hitchcock indicate predominant north-northeast wind direction in early postglacial time. Dunes overlying lake-bottom deposits and stream-terrace deposits in glacial Lake Hitchcock basin indicate predominant north-northwest wind direction after lake drainage; major dune fields cover lake-bottom surface on east side of valley. Dune sand now fixed by vegetation except where disturbed by human activities. Eolian silty sand, generally less than 1 m (3 ft) thick, is widespread in valleys and lower till slopes, but is not shown on map

ch **Submerged fluvial-estuarine, channel-fill deposits**—Fluvial sediments overlain by estuarine sediments (inferred from seismic-reflection data) up to 20 m (66 ft) thick in channel-fill configuration overlying steep-sided, channel-shaped unconformities that truncate glacial-lake deposits. Lower part of channel-fill sequence is complex; includes hummocky, lenticular, and short oblique clinoform reflectors suggesting cut-and-fill origin; interpreted from diatoms in vibracores (Szak, 1987) to be terrestrially derived fluvial sediment deposited when streams drained across subaerially exposed lakebed. Map pattern of these channels shows a paleodrainage system related to terrestrial valleys. Tributary channels draining southward from Connecticut and northward from Long Island join an east-draining trunk valley that has thalweg altitudes in the -40-m (-131-ft) range in the west and slopes to about -60 m (-197 ft) in the east where it exited the Long Island Sound basin at The Race. Fluvial facies is commonly overlain in the upper section of channel-fill by a parallel-laminated to seismically opaque unit interpreted to be fine-grained, estuarine sediment deposited as the rising postglacial sea entered the basin through -60-m (-197-ft) notch at The Race and spread to the west via a paleochannel system. Estuarine sediment extends outside the channel system in many places but is not mapped

Submerged marine deltaic deposits—Includes delta topset and foreset facies (inferred from seismic-reflection data) as much as 40 m (131 ft) thick that grades southwestward to delta-distal, fine-grained facies a few to 10 m (33 ft) thick, and associated paleoshoreline deposits. Marine deltaic sequence records a relative sea level of -40 m (-131 ft) in basin (Stone and Lewis, 1991). Contains an estimated 11.5 billion m^3 (15 billion yd^3) of material and is much thicker, coarser grained, and more internally complex than surrounding and overlying marine sediments (not shown on map). Sediment supply was via channel (ch) incised in lake beds that slopes from -22 m (-72 ft) at Connecticut River mouth to -35 m (-115 ft) south of Long Sand Shoal. During early stages of marine incursion, terrigenous sediment input via Connecticut River valley was low because glacial Lake Hitchcock in upper Connecticut River valley to the north persisted and formed a long-lived sediment trap. Large volume of sediment required to build marine deltaic sequence became available when glacial Lake Hitchcock drained at ~13.5 ka (Stone and Ashley, 1992) and newly formed Connecticut River north of breached dam rapidly incised bed of glacial Lake Hitchcock

md Deltaic facies—Internal reflectors within delta consist of long, southwest-dipping, oblique-tangential clinoforms interpreted as sandy delta-foreset facies and packages of chaotic reflectors interpreted as coarser grained beds (locally delta-topset facies). Delta-foreset facies generally occurs in prograded-fill configuration overlying a wave-cut unconformity, and is present between -60 m (-197 ft) and -42 m (-138 ft) below sea level. Delta-topset facies occurs as high as -30 m (-198 ft) in altitude; interpreted delta topset-foreset contact lies at about -42 m (-138 ft). Deltaic beds occupy eastern half of deposit; area east of dotted line has undergone intense modern tidal scour and only remnants of delta deposits remain

mdd Delta-distal facies—Thin, parallel-laminated internal reflectors interpreted as a delta-distal, fine-grained facies. Overlies wave-cut unconformity in an onlap-fill configuration to as high as about -40 m (-131 ft); relict shoreline features (beaches, bars, or spits) that lie at -42 m (-138 ft) in the southwest and -36 m (-118 ft) in the northwest are associated with outer edges of delta-distal facies. Continuous reflectors can be traced across delta-distal facies, indicating that these levels were isochronous; difference in altitude of paleoshoreline between north and south attributed to glacio-isostatic tilting. Top of delta-distal facies is cut by minor unconformity with up to 4 m (13 ft) of relief. Vibracores penetrating unconformity indicate delta-distal facies to be finely laminated, very fine sand

Glacial Meltwater Deposits (Late Wisconsinan)

Glacial meltwater deposits are sorted and stratified sediments composed of gravel, sand, silt, and clay, including minor lenses of flowtill and other diamict sediment, deposited by flowing glacial meltwater. Mineralogy of sediments is highly variable across the State, but in general is closely similar to subjacent and northerly adjacent bedrock (see figure 3 on sheet 2). Gravel clasts and sand grains are generally fresh and nonweathered.

Sedimentary Facies and Morphosequences

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

Map Units

Each of the 204 correlated map units of meltwater deposits on this map is a group of morphosequences deposited along the same or related paths of meltwater flow. Each unit was deposited either in a single glacial lake, a related series of lakes, or along meltwater streams in a valley where no ponding occurred. The position of groups of morphosequences (map units) in the landscape further indicates the systematic northward retreat of the ice margin. Where drainage divides were parallel or oblique to the trend of the ice margin, groups of high-level deltaic sediments were deposited when paths of meltwater escape were first held to higher positions against or through uplands, and then gradually lowered as lower paths were uncovered in valleys. On the basis of stratigraphic relationships between deposits, successive retreating ice-margin positions, and changes in glacial lake levels and in paths of meltwater flow, morphosequences are grouped into map units that are chronostratigraphic in character and that define a relative chronology of ice retreat across the State (see figure 4 on sheet 2).

Depositional Systems

Six depositional systems of meltwater deposits have been identified in Connecticut as a result of regional synthesis. On the map, units are grouped by color as follows:

Blues—Major ice-dammed lakes (IL)
Greens—Major sediment-dammed lakes (SL)
Purples—Related series of ice-dammed ponds (IP)
Browns—Related series of sediment-dammed ponds (SP)
Light oranges—Proximal meltwater streams (FP)
Dark oranges—Distal meltwater streams (FD)

Each depositional system is defined by lithostratigraphic principles and is characterized by morphosequence types, by spatial arrangements of sedimentary facies, and by typical stratigraphic relationships between individual deposits. The six depositional systems represent meltwater deposition in six paleogeographic settings that formed repeatedly in time and space, consequent to the interaction between the ice margin and the landscape over which it retreated.

Ponding of meltwater occurred in nearly every valley in the State during deglaciation; as a result, most meltwater sediments were deposited in or graded to glacial lakes. Four of the six depositional systems formed in paleogeographic settings in which lakes controlled the distribution and altitude of fluvial, deltaic, and lake-bottom sediments. Lakes controlled deposition in north-draining valleys, which sloped toward the retreating ice margin. In these valleys, the ice margin impounded meltwater against opposing topography and spillways were located across the lowest points of drainage divides. The resulting water bodies are referred to as ice-dammed glacial lakes and ponds. Lakes also controlled deposition in most south-draining valleys, which sloped away from the ice margin. In these valleys, lakes were impounded behind thick, valley-filling bodies of sediment that were constructed during preceding and successive meltwater deposition in each valley. The resulting water bodies are referred to as sediment-dammed lakes and ponds. Each sediment dam was itself an ice-marginal meltwater deposit, graded to a slightly older lake in the valley. Spillways for succeeding lakes commonly were over these dams.

Meltwater deposition in streams that were not tributary to glacial lakes was relatively uncommon in Connecticut. Glaciofluvial sediments were deposited in positions both proximal and distal to the ice margin.

Glaciolacustrine Systems

Sediments of four glaciolacustrine systems (IL, SL, IP, and SP) were deposited in or graded to glacial lakes and ponds. Sediments deposited in lakes include delta foreset and bottomset beds, lake-bottom sediments, and local lacustrine fan sediments. Lake-bottom sediments in all map units are shown by a horizontal line pattern. Sediments graded to glacial lakes include fluvial delta topset beds, delta-tributary fluvial sediments (fluvial sediment graded to deltas), and local ice-channel sediments. Delta-tributary fluvial sediments in large glacial-lake map units are shown by a dot pattern. Deposits of glaciolacustrine systems are predominantly deltaic. Altitudes of topset-foreset contacts in deltas record the paleo-water-plane altitudes of the glacial lake into which they were built. Deltas of all glacial lakes in Connecticut indicate paleo-water-plane slopes of 0.9 m/km (4.74 ft/mi) to the north-northwest. This slope is due to the glacio-isostatic tilt of the Earth's crust.

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

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

Glaciofluvial Systems

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

Uncorrelated Meltwater Deposits

u **Sand, gravel, silt, and clay**—Deposited in unidentified systems. Most appear to be ice-contact deposits entirely on glacial ice and collapsed down to present positions in landscape. Locally includes lake-bottom deposits (ruled pattern) that cannot be associated with any particular glacial lake

Deposits of Major Ice-Dammed Lakes—IL Depositional System

Paleogeographic setting: Gently sloping, relatively wide valleys and basins with drainage outlets to the north. Main valleys commonly fed by steeper tributary valleys. Lakes impounded in these valleys and basins when ice margin blocked drainage outlet to the north. Lakes spilled through cols floored in till and (or) bedrock across drainage divides. Some lakes had two or three stages as northward retreat uncovered lower spillways out of the basin.

Deposits: Deltaic, fluvial, and lake-bottom sedimentary facies are included in these deposits. Delta-tributary fluvial sediments shown by dot pattern; lake-bottom sediments shown by line pattern. Most prevalent morphosequences are ice-marginal deltas, but ice-marginal and near-ice-marginal fluviodeltaic sequences also occur; locally, ice-marginal lacustrine fan deposits are found. Lake-bottom deposits associated with multiple deltaic morphosequences cover large areas.

Stratigraphic arrangement of deposits: Deltas in this depositional system generally have free fronts that grade into flat-lying lake-bottom sediments that separate younger deltas from older ones. Altitudes of deltas reflect stable lake levels within each identified lake stage. Surfaces of deltas rise to the north only in the amount that they have been tilted by postglacial rebound. Lower groups of deltas graded to lower spillways occur in those lakes that have two or more stages identified. Fluvial facies within fluviodeltaic morphosequences occur predominantly in tributary valleys to the main basin or wide valley that contained these glacial lakes.

Map units: Shown in blue colors on the map. Unit descriptions (below) include information on location of the particular glacial lake and altitude of associated spillway, location and altitude of lower lake stages and associated spillways, altitudes of deltaic deposits, location of fluvial sediments in tributary valleys (if present), indication of buried lake-bottom sediments, and thicknesses (if known and of importance). Each map unit description also may include discussion of damming mechanism and ice-margin positions requisite to formation of the glacial lake. In some cases, geologic history is discussed.

Housatonic–Southwest Coastal Basin

ln **Glacial Lake Norfolk deposits**—Ice-marginal deltaic deposits and lake-bottom deposits in northwest-sloping Blackberry River valley. Deltas record lake levels in two stages. Deltaic deposits at Norfolk with surface altitude of 410 m (1,345 ft) were controlled by a spillway at 404 m (1,325 ft) across Housatonic and Naugatuck Rivers drainage divide. Ice-marginal deltaic deposits at East Canaan with surface altitudes at 270 m (885 ft) and topset-foreset contacts at about 265 m (870 ft) were controlled by 261-m (855-ft) spillway to northwest around northern end of Canaan Mountain. (Ashley Falls, South Sandisfield, Norfolk)

lcw **Glacial Lake Cornwall deposits**—Deposits in southern part of unit indicate either highly collapsed ice-marginal deltas or (more likely) ice-marginal lacustrine fans that did not build up to lake level. Surface altitudes are well below 337-m (1,105-ft) col across southern divide of Valley Brook. Northern deposits include small ice-marginal deltas built into lake from Baldwin Brook and Bloody Brook valleys at three levels: 261 m (855 ft), 230 m (755 ft), and 215 m (705 ft); highest level probably was controlled by 258-m (845-ft) spillway high on southern side of Furnace Brook valley, while ice margin blocked drainage. Lower two levels must have been controlled by ice or till blockage in Furnace Brook valley. Thick lake-bottom deposits occur beneath and in front of 215-m (705-ft) deltaic deposits in Cornwall village. (Cornwall)

lho **Glacial Lake Hollenbeck deposits**—Ice-marginal deltaic and lacustrine fan deposits and lake-bottom deposits in northwest-sloping Hollenbeck River valley. Successive ice-marginal deltas with narrow, ridge-like forms at southern end of lake are at 300 to 303 m (985 to 995 ft) and were controlled by 297-m (975-ft) spillway across divide at southern end of valley. To the north, successive ice-marginal lacustrine fan deposits are at lower than lake-level altitudes; in several exposures, internal bedding of these sediments displays severe, active ice deformation. Later lake stages were controlled by spillways at 215 m (705 ft) and 203 m (665 ft) west of Cobble Mountain. (South Canaan, Cornwall)

Glacial Lake Danbury deposits—Glacial Lake Danbury was dammed by northerly retreating ice margin in north-sloping Still River valley. Lake had three stages controlled in level by three successively lower spillways

ldh Pumpkin Hill stage deposits—One ice-marginal delta at 114 m (375 ft), controlled by short-lived 114-m (375-ft) spillway across northern end of Pumpkin Hill, adjacent to junction of Still River valley with Housatonic River valley. More extensive ice-marginal deltaic deposits at 90 to 93 m (295 to 305 ft) were graded to last stage of glacial Lake Danbury and controlled by 90-m (295-ft) spillway about 305 m (1,000 ft) farther north on Pumpkin Hill. Narrow bedrock gorge of Housatonic River just downstream from spillway must have been temporarily blocked to permit use of Pumpkin Hill spillway during ice retreat for about 3 km (2 mi) north of gorge. (New Milford)

ldp Pond Brook stage deposits—Ice-marginal and near-ice-marginal deltaic deposits, esker-fed lacustrine fan deposits, and associated lake-bottom sediments. Near-ice-marginal deltas south of Lake Candlewood have surface altitudes of 126 to 130 m (415 to 425 ft); inset against these is a delta at 114 to 117 m (375 to 385 ft) that was fed by distal meltwater from Lake Candlewood valley. Ice-marginal delta deposits just west of spillway are at 120 to 123 m (395 to 405 ft); other ice-marginal deposits are lacustrine fans that did not build up to lake level. This stage of glacial Lake Danbury was controlled by spillway eastward into Pond Brook valley that began initially at about 126 m (415 ft) and was incised down to about 114 m (375 ft) during time of its use. (Danbury, New Milford)

lds Saugatuck divide stage deposits—Ice-marginal deltas, fluviodeltaic deposits, and associated lake-bottom sediments. Delta surfaces have altitudes of 136 to 139 m (445 to 455 ft); fluvial feeders reach 154 m (505 ft). This stage of glacial Lake Danbury was controlled in level by 127-m (415-ft) spillway south into Saugatuck River valley. (Danbury, Bethel)

lpt **Glacial Lake Pootatuck deposits**—Successive ice-marginal deltaic deposits in north-draining Pootatuck River valley; near-ice-marginal fluviodeltaic deposits in upper valley. Deltas at 133 to 139 m (435 to 455 ft) in southern part of unit were graded to 130-m (425-ft) and 133-m (435-ft) spillways across southern divide. Ice-marginal deltas in Berkshire village at 117 to 120 m (385 to 395 ft) were graded to 114-m (375-ft) spillway to the east; northernmost ice-marginal deltas at 102 to 105 m (335 to 345 ft) were controlled by spillway across Berkshire deltas northward into Pole Bridge Brook valley; floor of this spillway is now at 96 m (315 ft) but was initially 3 to 6 m (10 to 20 ft) higher. (Newtown, Botsford)

Naugatuck Basin

lw **Glacial Lake Winsted deposits**—Ice-marginal deltas and fluviodeltaic deposits. Deltas have surface altitudes at 224 to 230 m (735 to 755 ft); extensive fluvial deposits in main valley reach 239 m (785 ft) and in tributary valleys are as high as 274 m (900 ft). Lake was ponded in north-draining Still River valley and controlled in level by three spillways. First was at 224 m (735 ft) and spilled into Naugatuck River valley. Next two successive spillways near Winsted drained into Farmington River valley. These spillways now are at 227 m (745 ft), but were 6 to 9 m (20 to 30 ft) lower relative to the first one before postglacial tilting. (Winsted, Torrington)

Farmington–Quinnipiac Basin

lnp **Glacial Lake Nepaug deposits**—Ice-marginal deltas in lower (eastern) part of Nepaug River valley and near-ice-marginal deltaic deposits in upper valley. Both sets of deltas record two lake levels. Deltas with surfaces at 215 to 224 m (705 to 735 ft) were graded to lake levels controlled first by 218-m (715-ft) and then 212-m (695-ft) spillway on till-blanketed hillside south of Nepaug Reservoir. Deltas with surfaces at 200 to 203 m (655 to 665 ft) are inset into higher level deltas in upper valley and easterly against higher deltas in lower valley; they were controlled by 191-m (625-ft) spillway downslope from first spillways. Glacial Lake Nepaug was contained in interlobate angle between Connecticut Valley ice lobe and western upland ice margin; valley ice-dammed lake and ice-marginal deltas in lower Nepaug River valley were built against it; upland ice margin provided meltwater and sediment to build deltaic deposits in upper valley. Lake lowered as valley ice retreated eastward off hillside south of present Nepaug Reservoir. (Collinsville, Torrington)

lbr **Glacial Lake Bristol deposits**—Ice-marginal deltaic and fluviodeltaic deposits in upper Pequabuck River valley and its northern tributaries. Deltas with surfaces at 197 to 200 m (645 to 655 ft) near Terryville controlled by 194-m (635-ft) spillway across Naugatuck River–Farmington River–Quinnipiac River divide. Fluvial feeder deposits as high as 248 m (815 ft) in upper Poland River valley built from upland ice-margin positions and graded to deltas at Terryville. Ice-marginal deltaic deposits in northeastern part of unit were built from ice positions to the northwest and were probably controlled by separate local spillway at 209 m (685 ft). Glacial Lake Bristol was contained in an interlobate angle between Connecticut Valley ice lobe and western upland ice margin. Pequabuck River valley was dammed to very high levels against drainage divide by thick valley lobe ice; ice-marginal deltas were built into lake from lobe ice position; at same time, fluviodeltaic deposits were built into lake from ice to northwest. (Bristol, Thomaston, Collinsville)

Upper Connecticut Basin

Glacial Lake Coginchaug deposits

lcgm Middletown stage deposits—Deposits in lower Coginchaug River and Sumner Brook valleys consist of small ice-marginal deltas, mostly collapsed; narrow noncollapsed fringes of these deltas occur at 81 m (265 ft), 69 m (225 ft), and 56 to 59 m (185 to 195 ft) built into successively lower levels of this stage, controlled by three spillways across rock ridge west of Sumner Brook at 75 m (245 ft), 59 m (195 ft), and 53 m (175 ft). Valley of Sumner Brook also was ponded, spilling out via two successive spillways east of valley at 59 m (195 ft) and 53 m (175 ft). Only a few small deposits at this level exist in Sumner Brook valley. Lake-bottom deposits in Sumner Brook may have accumulated, in part, in glacial Lake Middletown, which succeeded glacial Lake Coginchaug in this valley at lower level. (Middletown, Durham)

lcgd Durham stage deposits—In upper Coginchaug River valley, ice-marginal deltas, and ice-marginal and near-ice-marginal fluviodeltaic deposits built by meltwater from tributary valleys. Delta surfaces are at 87 m (285 ft) at southern end of basin and 93 m (305 ft) at northern end. All deposits except very earliest were graded to 84-m (275-ft) spillway across southern Coginchaug River drainage divide. Lake-bottom sediments occur in three areas; largest body, beneath Durham Meadows, may have accumulated mostly during Middletown stage. (Durham, Middletown)

lma **Glacial Lake Manchester deposits**—Predominantly ice-marginal deltaic deposits graded to long, narrow lake between Connecticut Valley ice lobe and upland till slope in south-central Manchester. Two stages of lake were controlled by spillways between bedrock hills; earlier stage with delta surfaces at 84 to 99 m (275 to 325 ft) was controlled by 81-m (265-ft) spillway; later stage delta surfaces are at 66 to 78 m (215 to 255 ft) controlled by 66-m (215-ft) spillway. (Manchester, Rockville)

lsb **Glacial Lake Salmon Brook deposits**—Ice-marginal deltaic and fluviodeltaic deposits graded to two stages of long narrow lake initially in Salmon Brook valley, southwestern end of which was dammed by Connecticut Valley ice lobe, and then to the north in southeastern part of city of Manchester. In Salmon Brook valley, fluvial surfaces as high as 96 m (315 ft) grade southwestward to deltas at 81 m (265 ft) controlled by 75-m (245-ft) spillway between two drumlin hills. Ice-marginal deltaic deposits in Manchester have surfaces at 99 m (325 ft) and were controlled by two spillways: one over older 96-m (315-ft) surface has present altitude of 87 m (285 ft), and the other at 87 m (285 ft) between bedrock hills north of Salmon Brook valley. (Glastonbury, Manchester)

lrb **Glacial Lake Roaring Brook deposits**—Ice-marginal deltaic and fluviodeltaic deposits graded to three stages of long, narrow lake in Roaring Brook valley, southwestern end of which was blocked by Connecticut Valley ice lobe. Ice-marginal deltaic deposits with 136- to 139-m (445- to 455-ft) surfaces on southeastern side of Roaring Brook valley, graded to 126-m (415-ft) spillway. Fluvial deposits that head west of Minnechaug Mountain, north of valley, are at 130 m (425 ft) and grade southwestward into valley to deltas with surfaces at 111 to 114 m (365 to 375 ft), controlled by 105-m (345-ft) spillway. Predominantly deltaic deposits with surfaces at 108 to 114 m (355 to 375 ft) west of Minnechaug Mountain were graded to 105-m (345-ft) spillway across older deposits; meltwater from this spillway entrenched and terraced older 111- to 123-m (365- to 405-ft) surfaces and constructed deltas at southern end of unit with surfaces at 87 to 93 m (285 to 305 ft), controlled by 87-m (285-ft) spillway. (Glastonbury, Rockville)

Lower Connecticut Basin

lc **Glacial Lake Colchester deposits**—Successive ice-marginal deltas with surface altitudes of 114 to 120 m (375 to 395 ft) deposited in earlier stage of lake in Nelkin Brook and Meadow Brook valleys. Spillway across Colchester deposits (lc) began at about 117 ft (385 ft) and eroded down to bedrock at 111 m (365 ft). Ice-marginal deltas built into later stage of lake in Judd Brook and Raymond Brook valleys have surfaces at 120 to 130 m (395 to 425 ft); controlled by spillway at 114 m (375 ft). Meltwater flow across spillway eroded last delta of older stage of lake, because older stage drained while younger stage still existed. (Colchester)

lex **Glacial Lake Essex deposits**—Four small ice-marginal deltas graded to four early-stage spillways across southern Falls River drainage divide; ice-marginal fluviodeltaic deposits in Falls River valley and northern tributaries. Fluvial feeder deposits reach 35 m (115 ft) and grade to deltaic surfaces in Centerbrook at 11 to 14 m (35 to 45 ft); controlled by spillway at southern part of Essex village at 11 m (35 ft). Early ice-marginal deltas probably built at nearly same time, while ice still occupied Falls River valley. When Falls River valley was uncovered, ice margin continued to block mouth of valley north of Essex village, creating main part of lake basin. (Essex, Deep River)

Quinebaug and Southeast Coastal Basin

lon **Glacial Lake Oneco deposits**—Successive ice-marginal deltaic deposits in Moosup River valley and ice-marginal and near-ice-marginal fluviodeltaic deposits in northern tributary valleys, some of which are in Rhode Island. A free-front delta at Oneco with lake-bottom sediments in front of it was built westward into lake fed by ice-marginal fluvial deposits farther upstream in Moosup River valley in Rhode Island. Ice-marginal fluvial deposits in Quanduck Brook valley are at 169 m (555 ft) and grade southward on steep gradient to 123- to 126-m (405- to 415-ft) deltas built into lake. Lake was dammed by ice margin in west-northwest-draining Moosup River valley and was controlled by three spillways. Highest spillway at 114 m (375 ft) is just over border in Rhode Island as are the deltas it controlled; a lower spillway that controlled most deposits in Connecticut crosses Quinebaug River-Pawcatuck River divide at 111 m (365 ft). A 108-m (355-ft) notch cut into a thick till hillside at last ice-margin position may have controlled level of lake for a short period of time before further ice retreat caused lake to empty. (Oneco)

lvo **Glacial Lake Voluntown deposits**—Successive ice-marginal deltas related to at least seven ice-margin positions. Several esker-fed lacustrine fan deposits that did not quite build up to lake level occur in central part of lake basin. Lake-bottom deposits exist at surface in few small areas that remained as open water. Lake existed in three stages, controlled by successive spillways at 123 m (405 ft), 111 m (365 ft), and 102 m (335 ft) across Quinebaug River–Pawcatuck River divide. (Voluntown)

lp **Glacial Lake Pachaug deposits**—Predominantly successive ice-marginal deltas in Pachaug River basin including some esker-fed deltas. Lake-bottom sediments underlie meltwater-terrace deposits in central part of basin and probably occur beneath Pachaug Pond as well. Ice-marginal fluviodeltaic deposits occur as one morphosequence in northeastern part of unit. Deposits are related to at least 10 ice-margin positions. Glacial Lake Pachaug existed in three successively lower stages controlled by spillways across divide on western side of basin. Earliest delta was built into highest stage controlled by 72-m (235-ft) spillway when ice margin was at southern end of basin. Several succeeding deltas were deposited in second stage of lake controlled by 69-m (225-ft) spillway. Last and longest-lived stage of lake controlled by 59-m (195-ft) spillway is recorded by deltas built at six successively northward ice-margin positions. Further ice retreat caused glacial Lake Pachaug to drain westward into Quinebaug valley and caused water levels to lower to altitude of glacial Lake Quinebaug. (Jewett City)

Deposits of Major Sediment-Dammed Lakes—SL Depositional System

Paleogeographic setting: Gently sloping, relatively wide valleys and basins that drained to the south, away from the ice margin. Relatively large glacial lakes formed in these valleys and basins behind thick sediment dams that filled narrower sections of the valleys. Dams most commonly were composed of ice-marginal meltwater sediments (usually deltaic) deposited at slightly earlier ice-margin positions in the valleys. Lakes developed in wider sections of valleys or in basins within valleys and were commonly fed by streams in tributary valleys to the lakes. Spillways for some lakes were over their sediment dams, in which case the lake-level lowered continuously during the life of the lake because the spillway was across easily erodible sand and gravel deposits. Other lakes had spillways with their base in bedrock across basin divides that were lower in altitude than the surface of the sediment dam blocking the valley; lake levels were stable throughout the life of these lakes.

Deposits: Deltaic, fluvial, and lake-bottom sediments are included in these deposits. Delta-tributary fluvial sediments shown by dot pattern; lake-bottom sediments shown by line pattern. Morphosequence types include ice-marginal deltaic, ice-marginal fluviodeltaic, and near-ice-marginal fluviodeltaic deposits; locally, ice-marginal lacustrine fan deposits occur; glacial Lake Hitchcock deposits include meteoric deltas. Lake-bottom deposits associated with multiple deltaic morphosequences cover large areas.

Stratigraphic arrangement of deposits: Deltas in this depositional system commonly have free fronts with lake-bottom sediments beyond that separate younger deltas from older ones. Altitudes of deltas in most of these deposits rise to the northwest at a rate less than the slope of postglacial tilt, a result of lowering spillway levels. Exceptions are deltas of glacial Lake Quinebaug and stable-level deltas of glacial Lake Hitchcock; these deltas rise to the northwest at a rate equal to the slope of postglacial tilt and reflect stable lake levels controlled by bedrock spillways. Fluvial facies within fluviodeltaic morphosequences occur either in valleys tributary to the main basin or in wide valleys that contained glacial lakes.

Map units: Shown in green colors on the map. Unit descriptions include information on location of particular glacial lake and location and type of spillway (that is, whether it was a lowering spillway with a sand and gravel base across the sediment dam, a lowering spillway with a till base across a basin divide, or a stable spillway with a bedrock base). Several lakes had more than one type of spillway during their existence. Descriptions also indicate altitudes of deltas, location of fluvial deposits in tributary valleys, and presence of buried lake-bottom sediments. Each map-unit description also includes discussion of damming mechanism and ice-marginal deposition requisite to the formation of the glacial lake. In some cases, geologic history is discussed.

Housatonic–Southwest Coastal Basin

lg **Glacial Lake Great Falls deposits**—Ice-marginal deltaic deposits and extensive lake-bottom sediments in Housatonic River valley from Great Falls northward into Massachusetts. Ice-marginal deltas are at 206 to 209 m (675 to 685 ft) near Canaan village. Lake-bottom deposits are as much as 30 to 61 m (100 to 200 ft) thick in overdeepened bedrock basin in easily erodible marble. Spillway was lip of modern Great Falls, initially perhaps 3 to 5 m (10 to 15 ft) higher than present 192-m (630-ft) level. Lake extended northward into Massachusetts. (Ashley Falls, South Canaan)

ll **Glacial Lake Lime Rock deposits**—Ice-marginal deltas with surfaces at 181 to 187 m (595 to 615 ft) and lake-bottom sediments built into small lake at junction of Salmon Creek valley with Housatonic River valley. Lake was dammed behind ice-contact head of Housatonic River deposits (hrn); spillway across Housatonic River was removed by postglacial incision of river. (South Canaan, Sharon)

lk **Glacial Lake Kenosia deposits**—Extensive lake-bottom deposits at lower end; delta rises west to ice-contact head just across State line in New York. Water from this lake eroded thick till along Still River and spilled into glacial Lake Danbury. (Brewster, Danbury)

lb **Glacial Lake Bantam deposits**—Ice-marginal and near-ice-marginal deltas with surfaces at 276 to 282 m (905 to 925 ft). Initial spillway at 276 m (905 ft) across divide at southern end of present Bantam Lake; later control of glacial lake level was across till surface at present lake outlet. Basin in which glacial lake existed and in which modern lake remains formed by deposition of extensive thick till in Litchfield drumlin field. (Litchfield)

lpg **Glacial Lake Pomperaug deposits**—Ice-marginal deltaic and fluviodeltaic deposits, near-ice-marginal fluviodeltaic deposits, and associated lake-bottom sediments. Successive ice-marginal deltaic and fluviodeltaic deposits in southern part of unit have surface altitudes of 78 to 81 m (255 to 265 ft). Initial ponding was behind Housatonic River deposits (hrs); spillway across those deposits was incised from about 72 m (235 ft) to 66 m (215 ft). Northern part of unit consists largely of fluviodeltaic deposits; long fluvial feeder deposits reach 154 m (505 ft) in Weekeepeemee River valley and 123 m (405 ft) in Nonewaug River and Sprain Brook valleys and grade to deltaic deposits at Woodbury with surface altitudes of 87 to 96 m (285 to 315 ft). This later lake drainage spilled across earlier deposits at Pomperaug village and may have finally stabilized in bedrock spillway at 78 m (255 ft). (Woodbury, Southbury)

Farmington–Quinnipiac Basin

lt **Glacial Lake Tariffville deposits**—Deltaic deposits overlying lake-bottom sediments more than 30 m (100 ft) thick in lower Salmon Brook valley. Ice-marginal fluvial deposits at Goodrichville reach 84 m (275 ft) and are on grade with deltaic surfaces to the south at 75 m (245 ft). West of Manitook Mountain, highly collapsed ice-marginal deposits grade to delta surfaces as high as 84 m (275 ft) that contain as much as 9 m (30 ft) of topset beds. Level of glacial Lake Tariffville was controlled at Farmington River gap through Talcott Mountain and was most likely water plane of glacial Lakes Middletown and Hitchcock to the east. Some slightly lower surfaces at 59 to 66 m (195 to 215 ft) are terraces or inset deltas graded to lowering water levels in glacial Lake Hitchcock. (Tariffville)

lf **Glacial Lake Farmington deposits**—Ice-marginal fluviodeltaic deposits in Roaring Brook valley near Unionville and west of Farmington River in Avon. Fluvial surfaces reach 93 m (305 ft) and are on grade southward to deltaic surfaces at 81 to 84 m (265 to 275 ft) at Unionville and 78 to 81 m (255 to 265 ft) in Farmington. Deposits east of Farmington River in Avon are successive ice-marginal deltas with several feeder eskers; deltaic surfaces are at 84 to 87 m (275 to 285 ft). South of Farmington River, fluviodeltaic deposits were built against tongue of ice in valley; deltaic surfaces are at 72 m (235 ft), and fronts of these deltas were erosionally trimmed by later meltwater that deposited Quinnipiac River valley terrace deposits (qt) in central portions of valley. Distal lacustrine sand containing climbing ripples that indicate northward current directions can be seen in several places beneath unit qt; this material was supplied to lake by meltwater from Pequabuck River valley carrying sediment derived from crystalline rock. Spillway for this lake was initially across till deposits incised from about 67 m (220 ft) to 58 m (190 ft); subsequent melting of buried ice in Quinnipiac River valley and lowering of glacial Lake Southington caused outlet to switch positions so that spillway was across deposits of glacial Lake Southington (ls). (New Britain, Bristol, Avon, Collinsville)

ls **Glacial Lake Southington deposits**—Successive ice-marginal esker-fed deltas in Quinnipiac River valley, each of which has a collapsed proximal head on northern side and free-front delta slopes on southern side. Lake-bottom sediments occur in front of each delta. In western tributary valley, single ice-marginal fluviodeltaic deposit occurs with lake-bottom sediments at southern end. Delta topset-foreset contacts in towns of Cheshire and Southington occur at 59 to 60 m (193 to 196 ft), and indicate spillway at about 56 m (185 ft) over blocking material near present Quinnipiac Gorge. Quinnipiac River–Mill River deposits (qm) occur at 58 m (190 ft) on both sides of Quinnipiac River immediately west of gorge and provided some blockage to dam the lake. Gorge probably was blocked with thick deposits of till; however, it is possible that bedrock gorge did not yet exist and was carved by water spilling from glacial Lake Southington and succeeding lakes to the north. (Southington, Meriden, Bristol, New Britain)

lq **Glacial Lake Quinnipiac deposits**—Consists of an extensive delta in North Haven and small area of lake-bottom sediments exposed by clay pit operations. Entire lower Quinnipiac River valley from north of Fair Haven to South Meriden contains lacustrine deposits of glacial Lake Quinnipiac in subsurface, beneath extensive Quinnipiac River valley fluvial terrace deposits (qt); lacustrine deposits consist principally of varved silt and clay (New Haven clay of Flint, 1933) as much as 49 m (160 ft) thick. Delta at North Haven has ice-margin position on its north side, but foreset dip directions indicate that it was built mainly from meltwater coming from the northeast down Muddy River valley. Topset-foreset contacts (Lougee, 1938) in this delta indicate a 9- to 11-m (30- to 35-ft) altitude water plane at Fair Haven where glacial Lake Quinnipiac spilled across lobe of New Haven delta plain (lcnh), which had built out across former Quinnipiac River channel and blocked valley. New Haven delta acted as sediment dam to impound glacial Lake Quinnipiac above level of glacial Lake Connecticut during construction of Muddy River delta. Water spilling from glacial Lake Quinnipiac probably fairly quickly incised dam to level of glacial Lake Connecticut to the south; however, no exposed deltas exist to the north of Muddy River delta to record lower lake levels. (Branford, New Haven, Wallingford)

Upper Connecticut Basin

Glacial Lake Hitchcock deposits—See discussion of lake in accompanying text

lhbe Beach deposits—Terrace-like deposits of sand on eastern shore of glacial Lake Hitchcock, ranging from 35 m (115 ft) in the south to 44 m (145 ft) in the north. These altitudes project to stable 25-m (82-ft) level of lake; some areas of beach deposits show slightly higher levels up-slope to the east. (Broad Brook, Manchester)

lhlb Lake-bottom deposits—Varved silts and clays in couplets ranging from 1 to 4 cm (0.4 to 1.6 in) thick and probably representing annual deposition. Well logs indicate that silt-clay couplets are thicker and red in color in lower part of section; these are overlain by alternating red and gray sediments; silt-clay couplets are thinnest in upper part of section and are gray. Numerous surface exposures indicate that top 2 to 3 m (6 to 10 ft) of section consists of thinly bedded silt and fine sand of sandy lake-bottom facies. Deposits are areally extensive and commonly greater than 15 m (50 ft) thick, as much as 80 m (260 ft) thick in some places

Glacial Lake Hitchcock, high-level deposits—Ice-marginal deltas and fluviodeltas, distal-meltwater-fed deltas, and a meteoric-water-fed delta. Delta altitudes adjusted for glacio-isostatic tilting and projected to New Britain spillway (NBS) record early lake levels from 35 m (115 ft) to about 29 m (95 ft)

lhhw Windsor deltaic deposits—Ice-marginal deltaic deposits in northeastern part of unit with surfaces at 59 to 61 m (195 to 200 ft); topset-foreset contact at 54 m (178 ft) (projects to NBS at 35 m (114 ft)). Deltaic deposits in southwestern part of unit with surfaces at 50 to 56 m (165 to 185 ft) were fed by distal meltwater; sediment was supplied to lake by early Farmington River–Salmon Brook meltwater drainage eastward through Tariffville Gap. (Windsor Locks, Hartford North)

lhhe East Windsor fluviodeltaic deposits—Highly collapsed fluvial deposits with remnant surfaces as high as 69 to 72 m (225 to 235 ft) in northern part of unit grade southward along ice-margin position to deltaic deposits at 53 to 56 m (175 to 185 ft) in southern part of unit. (Broad Brook)

lhhh High-level Hockanum River delta deposits—Small delta with surface at 38 to 41 m (125 to 135 ft) and topset-foreset contact about 35 m (115 ft) (projects to NBS at 32 m (104 ft)); built by meteoric water in Hockanum River and Hop Brook valleys. (Manchester)

lhhs High-level Scantic River delta deposits—Ice-marginal delta with surface at 59 to 62 m (195 to 205 ft) in southern part of unit. Distal-meltwater-fed delta with surfaces at 56 to 59 m (185 to 195 ft) in northern part of unit. (Broad Brook)

lhhr Rattlesnake Brook deltaic deposits—Ice-marginal deltas with surfaces at 59 to 62 m (195 to 205 ft). (West Springfield, Windsor Locks)

lhhsc Shea Corner deltaic deposits—Ice-marginal deltaic deposits that extend into Massachusetts. Surfaces at 59 to 62 m (195 to 205 ft). (West Springfield)

lhhrn Enfield deltaic deposits—Deltaic deposits with surfaces at 59 to 62 m (195 to 205 ft) are distal parts of fluviodeltaic deposits built from ice-margin position north of map area in Massachusetts. Estimated lake level of 58 m (190 ft) (projects to NBS at 30 m (99 ft)). (Springfield South)

lhhi Ice-hole deposits—Collapsed, ice-contact sand and gravel deposits that encircle hills (mostly drumlins) which were islands in glacial Lake Hitchcock. As ice margin in lake thinned and melted away from island hillsides, deposits probably fed by subglacial meltwater streams accumulated in ice-walled depressions around drumlin hillsides. Glacial Lake Hitchcock water levels apparently controlled altitude of these deposits, because none were built to higher levels. (West Springfield, Hartford North, Manchester)

lhhp Spit deposits—Northeast-trending spits composed of southwest-dipping sandy foreset beds on southwestern ends of drumlin islands; generally smaller in size on drumlins lacking ice-hole deposit collar and larger in size on those that have it. Spits were built by waves and currents generated by early paleowinds from the northeast when ice margin was still nearby. (West Springfield, Windsor Locks, Broad Brook, Hartford North, Manchester)

Glacial Lake Hitchcock, stable-level deposits—Meteoric-water-fed deltas and extensive lake-bottom deposits. Delta altitudes, when adjusted for glacio-isostatic tilting and projected to NBS, record 25-m (82-ft) stable lake level that began when ice margin was in southern Massachusetts

lhsb Bradley International Airport delta deposits—Large delta built by meteoric water from Farmington River valley that flowed through Tariffville Gap and discharged northeastward into glacial Lake Hitchcock. Delta probably is complex; in places in subsurface, older foreset beds built by distal meltwater from Farmington River valley into higher lake levels may lie beneath present delta surface. Delta surface slopes from 56 m (185 ft) in southwest to 47 m (155 ft) in the north and east. Topset-foreset contact at 47 m (154 ft) (projects to NBS at 25 m (82 ft)). (Windsor Locks)

lhss Stable-level Scantic River and Broad Brook delta deposits—Small deltas with surfaces at 47 to 50 m (155 to 165 ft) built by meteoric water in Scantic River and Broad Brook valleys as it entrenched high-level deposits and entered lower lake level. (Broad Brook)

lhsh Stable-level Hockanum River delta deposits—Small delta with surface at 32 to 35 m (105 to 115 ft) built by meteoric water in Hockanum River and Hop Brook valleys as it entrenched deposits of the high-level Hockanum River delta deposits (lhhh) and entered lower level lake. (Manchester)

lhf **Glacial Lake Hitchcock post-stable-level Farmington River delta deposits**—Meteoric-water-fed delta with surface altitude of 41 m (135 ft) and topset-foreset contact at 39 m (127 ft); this altitude projects to NBS at 21 m (70 ft), which is 4 m (12 ft) lower than stable level. This delta represents relatively short period of time during which level of glacial Lake Hitchcock was lowered, probably due to first effects of postglacial uplift and before lake drained due to failure of dam. Farmington River terrace deposits (ft) are fluvial deposits associated with this delta; these terraces are inset into stable-level delta surface of Bradley International Airport delta deposits (lhsb). (Windsor Locks, Hartford North)

Glacial Lake Middletown deposits—Ice-marginal deltas and fluviodeltaic deposits that record slowly lowering lake level (when adjusted for glacio-isostatic tilting) as ice margin retreated from Portland, Middletown, and Berlin northward to area near Windsor and East Windsor. See discussion of lake in accompanying text

lmg Great Pond delta deposits—Ice-marginal delta with surface altitude of 59 to 62 m (195 to 205 ft); topset-foreset contact at about 55 m (182 ft) (projects to spillway at 21 m (68 ft)). (Windsor Locks)

lmwv Windsorville deltaic deposits—Ice-marginal deltaic deposits at 59 to 62 m (195 to 205 ft) built at the same time as Great Pond delta deposits (lmg) on western side of ice lobe; ice-proximal side of these deposits subjected to slight readvance of ice margin, which pushed deltaic material up into a small moraine (m). (Broad Brook)

lmw Western margin deltaic deposits—Ice-marginal deltas in New Britain and West Hartford with surface altitudes of 50 to 53 m (165 to 175 ft). Fluviodeltaic deposit at southern edge of New Britain has deltaic surface at 50 m (165 ft) and was built by meltwater flowing eastward through Cooks Gap. (Avon, New Britain)

lme Eastern margin deltaic deposits—Ice-marginal fluviodeltaic deposits; fluvial deposits as high as 78 m (255 ft) just west of Rockville in Hockanum River valley grade southwestward to deltaic surfaces near Manchester at 53 to 56 m (175 to 185 ft). Distal deltaic sands and lake-bottom sediments are highly collapsed south of Manchester. (Ellington, Rockville, Manchester, Glastonbury)

lmh Hockanum River delta deposits—Meltwater-fed delta with surface altitude of 47 m (155 ft) and topset-foreset contact at 42 m (138 ft) (projects to spillway at 20 m (65 ft)). Constructed by meltwater spilling from glacial Lake Ellington, entrenching and reworking deposits of eastern margin deltaic deposits (lme) as meltwater entered glacial Lake Middletown at southern end of Hockanum River valley. Inset fluvial terrace deposits (ht) in Hockanum River valley are on grade to delta. (Manchester)

lmn Newington deltaic deposits—Highly collapsed ice-marginal delta deposits. Some noncollapsed delta surfaces remain at 50 m (165 ft) in southern part of unit and at 53 m (175 ft) in northern part. (Hartford South)

lmc Cromwell deltaic deposits—Extensive ice-marginal deltaic deposits with surfaces at 44 to 47 m (145 to 155 ft) and topset-foreset contact at 41 m (135 ft) on western side of Connecticut River (projects to spillway at 33 m (107 ft)); deltas on eastern side of river, built slightly earlier, are at 50 to 56 m (165 to 185 ft) and have topset-foreset contacts at 45 m (149 ft) (projects to spillway at 37 m (120 ft)). Deltas in Cromwell have free fronts built into open water in glacial Lake Middletown basin. (Middletown, Hartford South, Glastonbury)

lmp Portland deltaic deposits—Successive ice-marginal deltaic deposits with surface altitudes of 47 to 50 m (155 to 165 ft); topset-foreset contacts exposed in several sand and gravel pits were measured at altitudes between 43 and 46 m (140 and 150 ft) (when projected, these reflect lowering at the spillway from 40 m (130 ft) to 37 m (120 ft)). (Middle Haddam)

lmb Lake-bottom deposits—Reddish brown, mostly varved silt and clay in couplets generally ranging from 5 to 10 cm (2 to 4 in) thick. Thickness and color of varves reflect annual deposition of beds while ice margin was in lake basin. Extensive deposits occur in basins of glacial Lakes Middletown and Berlin where section is up to 23 m (75 ft) thick

lso **Glacial Lake Somers deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits built along eastern margin of Connecticut Valley ice lobe as it lay southeast of present Scantic River in Somers. Deposits are highly collapsed, especially on northwestern sides; noncollapsed parts in earlier morphosequences have deltaic surfaces at 87 to 90 m (285 to 295 ft) and fluvial surfaces that reach 111 m (365 ft). Later morphosequences have deltaic surfaces at 78 to 81 m (255 to 265 ft). Surfaces north of North Somers at 90 to 93 m (295 to 305 ft) may be from ice-hole deltas. Lake-bottom sediments occur at surface in front of 78-m (255-ft) deltas in southern part of unit. Water levels were controlled by spillway over ice-contact head of glacial Lake Ellington deposits (le). Spillway was eroded from about 87 m (285 ft) to 72 m (235 ft) during life of lake. (Hampden, Ellington) le **Glacial Lake Ellington deposits**—Ice-marginal deltas with noncollapsed surfaces at 81 to 84 m (265 to 275 ft) near Sadds Mill and 84 to 87 m (275 to 285 ft) at and south of Ellington-Somers town line. Deposits grade southward to distal deltaic sand at Ellington village and lake-bottom sand and silt south of there. Spillway is not preserved but must have been over eastern margin delta deposits (lme) at about 72 m (235 ft). Meltwater discharging from this spillway cut terraces (ht) into glacial Lake Manchester (lma) and unit lme deposits farther south along Hockanum River valley, and built delta into glacial Lake Middletown (lmh). (Ellington)

Thames Basin

lu **Glacial Lake Uncasville deposits**—Ice-marginal deltas and fluviodeltaic deposits with deltaic surfaces at 17 to 20 m (55 to 65 ft) along Thames River and Poquetanuck Cove. Fluvial deposits on grade with deltas reach 29 m (95 ft) in tributary valleys. Coarse gravel fluvial deposits in lower Oxoboxo Brook valley reach 72 m (235 ft) and have very steep surface gradient to 20-m (65-ft) delta at Uncasville. Lake was ponded behind Ledyard moraine (hlm) and Jordan Cove–lower Thames River deposits (lcjl) in Thames River valley; sediment dam has largely been removed by postglacial Thames River entrenchment. (Uncasville, Montville, Norwich)

Quinebaug and Southeast Coastal Basin

lqb **Glacial Lake Quinebaug deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits and associated lake-bottom sediments. Delta-surface altitudes rise to the north from 47 m (155 ft) to 66 m (215 ft) due to glacio-isostatic tilting. Fluvial feeder deposits in tributary valleys reach as high as 87 m (285 ft). All deposits graded to single, stable lake level controlled by 41-m (134-ft) spillway over bedrock in narrow gorge where Quinebaug River turns westward, about 5 km (3 mi) south of Jewett City. Gorge probably was blocked by slightly older meltwater deposits (possibly by till); this led to establishment of drainage across adjacent 41-m (134-ft) bedrock saddle, which is about 21 m (70 ft) above present river level. Deposits of glacial Lake Quinebaug south of Quinebaug valley deposits (qb) are related to at least nine successive ice-marginal positions. Later deposits of unit lqb, north of unit qb in Blackwell Brook valley near Brooklyn, represent two successive ice-marginal deltas and associated lake-bottom deposits; altitudes of these deltas project to earlier glacial Lake Quinebaug water plane, probably due to erosion or collapse of unit qb deposits. (Jewett City, Plainfield, Danielson)

Long Island Sound Basin

Glacial Lake Connecticut deposits—Includes both onland and offshore deposits; see discussion of lake in accompanying text. Spillway for glacial Lake Connecticut at "The Race" is abbreviated below as TRS. All onland deltas have offshore extensions, which are included in offshore submerged deltaic deposit (lcd)

lcsnw Stamford–Norwalk–Westport deposits—Ice-marginal fluviodeltaic deposits in lower reaches of Rippowam, Noroton, Fivemile, Norwalk, and Saugatuck Rivers. Consist of fluvial sediments in the valleys of Rippowam, Noroton, and Fivemile Rivers; deltaic deposits (represented by lcd) are entirely submerged offshore from Stamford and Darien. Fluvial sediments reach altitudes of 20 m (65 ft) in Norwalk River valley and 26 m (85 ft) in Saugatuck valley and grade southward to deltaic surfaces at 5 to 8 m (15 to 25 ft) at Norwalk and Westport; bulk of these deltas, represented by lcd, are submerged below present sea level. (Stamford, Norwalk South, Sherwood Point, Westport)

lcss Stratford–Southport deposits—Fluvial deposits in Sasco Brook valley (as high as 26 m (85 ft)) grade to delta just west of Southport at 5 to 8 m (15 to 25 ft). Very coarse grained fluvial deposits in Mill River valley as high as 66 m (215 ft) grade southward to sandy delta in Fairfield at 5 to 8 m (15 to 25 ft). Fluvial deposits in Pequonnock River valley reach 26 m (85 ft) and grade to delta surface in Bridgeport at 5 to 11 m (15 to 35 ft). Ice-marginal fluvial deposits at 20 m (65 ft) in lower Housatonic River valley grade to delta in Stratford at 5 to 8 m (15 to 25 ft); these delta levels project to TRS at -14 to -15 m (-46 to -49 ft). (Westport, Bridgeport, Milford)

lcl Lordship deposits—Ice-marginal delta with ice-contact slopes on both northeastern and northwestern sides. Surface altitude of delta is 8 to 11 m (25 to 35 ft); topset-foreset contact estimated at 5 m (15 ft) (Hokans, 1952); -12 m (-39 ft) projected to TRS. Built slightly earlier into somewhat higher lake level than Stratford–Southport deposits (lcss) and Devon–Milford deposits (lcdm) to the north. Delta is likely contemporaneous with large, submerged, ice-marginal lacustrine fan deposit that extends eastward from Lordship (lcf). Ice-margin position recorded by this delta marks interlobate angle between western Connecticut ice margin and Connecticut Valley ice lobe. Till overlying deltaic sediments on northeastern side provides some evidence of slight readvance of Connecticut Valley ice lobe from the northeast. (Bridgeport, Milford)

lcdm Devon–Milford deposits—Ice-marginal fluvial deposits as high as 32 m (105 ft) in Beaver Brook valley grade southward to probable delta surface at 14 m (45 ft); distal end of this deposit was eroded by later meltwater in Housatonic River valley. Fluvial deposits in Wepawaug River valley with ice-contact head at 41 m (135 ft) and in Indian River valley at 32 m (105 ft) grade to delta surface in Milford at 8 to 11 m (25 to 35 ft); -15 m (-49 ft) projected to TRS. Ice-marginal deltaic deposits near Woodmont are at 8 to 11 m (25 to 35 ft). All deposits in this unit were built from ice-margin positions on western side of northeasterly retreating Connecticut Valley ice lobe. (Milford, Ansonia)

lcnh New Haven deposits—Fluvial deposits in West River valley as high as 47 m (155 ft) and ice-marginal fluvial deposits in Mill River valley as high as 35 m (115 ft) grade southward to massive delta plain at 11 to 17 m (35 to 55 ft) in New Haven, West Haven, and Fair Haven. Topset-foreset contacts of delta, seen in excavation for Yale Bowl in West Haven and along railroad cut in Fair Haven, are at about 9 m (30 ft) and 7 m (22 ft), respectively; these levels project to TRS at -18 to -20 m (-59 to -66 ft). Deltaic sediments in New Haven overlie more than 61 m (200 ft) of lake-bottom sediment. (New Haven, Mount Carmel)

lcenh East Haven deposits—Ice-marginal fluvial deposits in Farm River valley with ice-contact head at 32 m (105 ft) grade southwest to delta at East Haven with surface at 5 to 8 m (15 to 25 ft); -20 m (-66 ft) projected to TRS. (Branford, New Haven)

lcew East River–West River deposits—Ice-marginal fluvial sediments at 23 m (75 ft) in West River valley grade southward to delta at Guilford with surface altitude of 5 to 8 m (15 to 25 ft); -16 m (-52 ft) projected to TRS. Ice-marginal fluvial sediments reach 17 m (55 ft) in East River and Neck River valleys and grade southward to deltas at Madison, north of Madison moraine (mom) and head of Hammonasset River–Menunketesuck River deposits (lchm), with surface altitudes at 5 to 11 m (25 to 35 ft). (Guilford)

lchm Hammonasset River–Menunketesuck River deposits—Successive ice-marginal fluviodeltaic deposits. Fluvial sediments as high as 26 m (85 ft) in Menunketesuck River valley, 32 m (105 ft) in Indian River valley, and 17 m (55 ft) in Hammonasset River valley grade southward to deltas at Clinton with surface altitudes at 5 to 8 m (15 to 25 ft); -14 m (-46 ft) projected to TRS. Ice-marginal delta at Madison built along Madison moraine (mom) position has surface altitude of 5 m (15 ft); -14 m (-46 ft) projected to the spillway. Represents ice-marginal meltwater deposition graded to glacial Lake Connecticut following Hammonasset moraine (hlm) position and ending at Madison moraine (mom) position. (Clinton, Guilford, Essex)

lcwoo Westbrook–Old Saybrook–Old Lyme deposits—Successive ice-marginal deltaic and fluviodeltaic deposits built at mouth of Connecticut River, in its tributary valleys of Black Hall and Lieutenant Rivers, and along present shoreline in Westbrook. Consist of deltas built at three major ice-margin positions. Ice-marginal deltas with surface altitudes of 5 to 8 m (15 to 25 ft; -12 m (-39 ft) projected to TRS) were built into glacial Lake Connecticut from Old Saybrook moraine (owm) position. Ice-marginal delta north of South Cove in Old Saybrook and fluviodeltaic deposits in Black Hall River valley in Old Lyme at 5 to 8 m (15 to 25 ft) were built from second ice position. Ice-marginal deltas in Westbrook and Old Saybrook and fluviodeltaic deposits in Lieutenant River valley in Old Lyme were built from Hammonasset moraine (hlm) position. These deltas are at 5 to 11 m (15 to 35 ft) (-13 to -15 m (-43 to -49 ft) projected to TRS). (Essex, Old Lyme)

lcn Niantic deposits—Ice-marginal delta at mouth of Niantic River and ice-marginal fluviodeltaic deposits in Pataguanset River valley in East Lyme and in valleys of Threemile and Fourmile Rivers in Old Lyme were built from Old Saybrook moraine (owm) position; delta surfaces are at 8 to 11 m (25 to 35 ft), and well logs in Niantic village indicate topset-foreset contact is at about 1.2 m (4 ft) (-12 m (-39 ft) projected to TRS). In Bride Brook valley in East Lyme, coarse gravel fluvial deposits slope from 32 m (105 ft) with steep gradient to deltaic deposits at 8 to 11 m (25 to 35 ft) behind Old Saybrook moraine (owm) position; delta appears to be at glacial Lake Connecticut level, indicating that moraine did not exclude lake waters from mouth of valley. (Niantic, Old Lyme) lcjl Jordan Cove–Lower Thames River deposits—Ice-marginal fluviodeltaic deposits in Jordan Brook valley and several short tributaries to Long Island Sound were built from Old Saybrook moraine (owm) position. Scattered deltaic deposits along Thames River south of Ledyard moraine (hlm) position have surfaces at 8 to 14 m (25 to 45 ft); these levels are consistent with those projected for glacial Lake Connecticut. Only remnants of these deposits remain due to extensive dissection by later meltwater as level of glacial Lake Connecticut lowered; northern extent of large Thames River drainage basin contained distal meltwater flow for much longer period of time than did its smaller adjacent valleys such as Poquonock River. (Niantic, New London, Uncasville)

lcp Poquonock River deposits—Ice-marginal fluviodeltaic deposits in Poquonock River valley; fluvial sediments as high as 32 m (105 ft) in upper valley and small tributaries grade southward to extensive delta with surface altitudes of 5 to 11 m (15 to 35 ft) south of Interstate Route 95 in Groton. Well logs indicate possible delta topset-foreset contact at 1 to 2 m (2 to 5 ft) altitude (-11 to -12 m (-36 to -39 ft) projected to TRS). (Uncasville, New London)

lcms Mystic–Stonington deposits—Ice-marginal fluviodeltaic deposits in Mystic River, Copps Brook, Stony Brook, and lower Anguilla River valleys. Ice-marginal delta in lower Mystic River valley and fluviodeltaic deposits in small valleys to the east were built in association with Mystic moraine (mm) position; deltas have surfaces at 2 to 5 m (5 to 15 ft) and are mostly submerged below present sea level. Ice-marginal fluviodeltaic deposits in Mystic River valley head at Old Saybrook-Wolf Rocks moraine (owm) position. Fluvial sediments are at 20 m (65 ft) and grade to deltaic surfaces at 2 to 8 m (5 to 25 ft) in lower valley. Topset-foreset contact appears from well logs to be at sea level; this altitude projects to TRS at -9 to -10 m (-30 to -33 ft). (Mystic, Old Mystic)

Offshore submerged deposits of glacial Lake Connecticut

lclb Lake-bottom deposits—Inferred from seismic-reflection data to be varved silt and clay commonly 80 m (262 ft) thick and locally greater than 150 m (492 ft) thick in deep valleys. These deposits dominate glacial section in southern half of basin and variously overlie bedrock and (or) Cretaceous beds, undifferentiated drift, end-moraine deposits, and lacustrine-fan deposits. Unit is characterized by finely laminated, parallel internal reflectors that distinctively drape underlying topography. Several vibracores penetrated varved clay lake-bottom facies. One core (LISAT 6) contained 6.5 m (21 ft) of typical glaciolacustrine varved sediment in silt-clay couplets, which range from 0.7 to 7.1 cm (0.5 to 3 in) thick with a mean thickness of 2.2 cm (1 in) and (if interpreted as annual) represent 280 years of lacustrine deposition in interval sampled. Seismic-reflection data collected at core location reveal that another 30 m (98 ft) of lake-bottom clay facies is present in section beneath cored interval, and that local tidal scour has removed about 20 m (66 ft) of lake-bottom clay that formerly existed above cored interval. In many places, reflectors within lake-bottom clay facies can be traced northward into deltaic facies

lcd Deltaic deposits—Inferred from seismic-reflection data to be delta-foreset and bottomset facies of emergent deltaic deposits of coastal Connecticut; deposits are up to 40 m (131 ft) thick and are dominant component of lake sediment in much of northern nearshore area. Deposits locally overlie bedrock, undifferentiated drift, end-moraine deposits, or lacustrine-fan deposits; generally, however, they overlie and intertongue distally with varved clay lake-bottom facies (lclb). Internally, delta facies exhibits seaward-dipping, oblique-tangential reflectors; these progradational clinoform configurations occur most commonly on north-south-trending profiles. On east-west profiles, reflectors within delta facies are typically parallel to subparallel, horizontally stratified in fill configuration with lows in basal bounding surface. In several places between Old Saybrook and Clinton, deltaic facies reflectors can be traced northward directly into end-moraine deposits, indicating synchronous deposition of the two

lcf Ice-marginal lacustrine fan deposits—Present in lower part of glaciolacustrine section. Overlie bedrock, Cretaceous strata, and (or) undifferentiated drift and are commonly in same stratigraphic position as moraine deposits. Fans occur locally throughout basin, but are numerous and more extensive in wide central Long Island Sound. Each lacustrine-fan sequence consists of two facies that have different seismic characteristics; ice-proximal facies always occurs in northern part of deposit, and distal facies always occurs in southern part. On the map, each facies is distinguished by its own pattern, but both with the lcf label. Proximal facies is commonly an asymmetric, positive relief form with steeper southern slope and gentler northern slope. Inferred from seismic-reflection data to consist of coarse-grained sand and gravel in south-dipping beds; probably contains boulders (at least on the surface) and ablation till in most proximal parts. Deposits are typically either seismically amorphous or display chaotic internal reflectors; however, multiple seismic-reflection profiles run in different directions across single fans reveal that internal reflectors, which have chaotic configuration on some crossings, appear as steeply dipping clinoforms on other crossings. The seismic data provide evidence that proximal facies contains coarse-grained stratified sediment as well as nonstratified ablation material. Surface of distal facies slopes gently southward from higher-standing proximal facies. Inferred from seismic-reflection data to consist of finer grained lacustrine beds similar to overlying lake clays but deposited by turbidity underflow processes; characterized by finely laminated, subparallel to parallel internal reflectors that fill underlying topographic lows. Lacustrine fans were built beneath waters of glacial Lake Connecticut by meltwater streams that issued from grounding line of ice sheet. On several seismic-reflection profiles, evidence of systematic northward retreat of ice is provided by shingled sequences of up to 10 ice-marginal fan deposits. In these sequences, proximal facies of one fan is overlain by distal facies of next younger fan to the north

Deposits of Related Series of Ice-Dammed Ponds—IP Depositional System

Paleogeographic setting: Steeper, small valleys that slope to the north toward ice margin. Series of small lakes or ponds, impounded to the north by ice margin in one or several north-sloping valleys. Multiple spillways cut into till or bedrock across divides are at successively lower altitudes to the north. A few units formed in this depositional setting were built in a single north-draining valley, into only one small lake; most, however, formed in a series of lakes, which lowered successively to the north in several valleys descending from particular major or minor divides. Each group of ponds formed during retreat of the ice margin from impingement against the divide, and before uncovering of lower drainage outlets.

Deposits: Predominantly deltaic sedimentary facies are included in these deposits. Lake-bottom facies occur locally beneath the deltas, but are not exposed at the surface. Ice-marginal deltas are the only type of morphosequences present. Some ice-marginal deltas have fluvial feeder eskers (shown by chevron pattern).

Stratigraphic arrangement of deposits: Deltas generally do not have free fronts because the delta commonly completely filled the small body of water into which it was built and distal sand extended into the spillway entrance. Lake-bottom sediment consists mostly of bottomset beds (rather than extensive varved silts and clays) and occurs only beneath deltaic sediments. In any one valley, where deltas were built into lowering series of ponds, deltas commonly are contiguous; each younger delta surface laps against ice-contact proximal slope of the older delta.

Map units: Shown in purple colors on the map. Unit descriptions below include information on location of valley or valleys in which small glacial lake or series of lakes existed; spillway altitude, or range of altitudes of multiple spillways; and particular divide or divides that spillways crossed. Altitudes of delta surfaces (not usually given) are typically 2 to 3 m (5 to 10 ft) higher than altitude of related spillway due to thickness of fluvial topset beds. In some cases, geologic history is discussed.

ip **Uncorrelated deposits of ice-dammed pond**s

Housatonic–Southwest Coastal Basin

co **Cobble Brook deposits**—Four successive ice-marginal deltaic deposits in and adjacent to small valley on eastern side of Housatonic River. Deltas are mostly graded to 175-m (575-ft) spillway at head of valley. Low-lying deposits well below level of spillway were extensively collapsed. (Ellsworth, Kent)

ba **Bantam River deposits**—Series of ice-marginal deltas along Bantam River valley. Controlled by spillways at 239 m (785 ft) and 236 m (775 ft) at southwestern end and perhaps by some temporary higher spillways along edge of ice. Represents ice margin parallel to that of Mallory Brook deposits (ml), lower in Bantam River drainage so that numerous higher spillways of unit ml could no longer be used. (New Preston, Litchfield)

ml **Mallory Brook deposits**—Series of ice-marginal deltas along southeastern sides of Bantam River and Mallory Brook valleys. Controlled by six spillways ranging from 306 m (1,005 ft) down to 181 m (595 ft). (New Preston, Litchfield)

wd **Woodbury deposits**—Ice-marginal deltas in uplands east of Pomperaug River. Controlled by 236-m (775-ft) to 206-m (675-ft) spillways. Immediately succeeded by ice-marginal deltas perched on eastern side of river valley, controlled by 163-m (535-ft) to 102-m (335-ft) spillways. (Woodbury)

su **Southbury deposits**—Ice-marginal deltas in northwest-sloping valley on eastern side of Pomperaug River valley. Ponding initially controlled by 142-m (465-ft) spillway across Eightmile Brook deposits (eb) to the southeast. Successive levels controlled by three side-valley spillways to the southwest at 120 m (395 ft), 108 m (355 ft), and 102 m (335 ft). (Southbury)

sd **Still River–Saugatuck River divide deposits**—Ice-marginal deltas in three valleys that slope northward from Still River–Saugatuck River drainage divide. Level of ponding in each valley controlled by spillway across divide. Successive deltas in Limekiln Brook valley were controlled by 169-m (555-ft) spillway; one delta in Wolf Pit Brook valley was controlled by 197-m (645-ft) spillway; successive deltas in Sympaug Brook valley were controlled by 126-m (415-ft) spillway that also served as glacial Lake Danbury–Saugatuck River divide stage outlet. (Danbury, Bethel, Newtown, Botsford)

pd **Pootatuck River–Pequonnock River divide deposits**—Ice-marginal deltas in several valleys (including Pootatuck River) that slope northward from divide. Levels of ponding in Hurds Brook valley controlled by successively lower spillways at 157 m (515 ft), 154 m (505 ft), 136 m (445 ft), and 126 m (415 ft). Levels of ponding in two small tributary valleys to Halfway River controlled by two spillways at 126 m (415 ft) and 117 m (385 ft). Small pockets in upper reaches of Pootatuck River basin controlled by local spillways across divide at 157 m (515 ft) to 130 m (425 ft). (Newtown, Botsford, Long Hill)

Naugatuck Basin

to **Tolles–Terryville deposits**—Ice-marginal deltas ponded in small upland valleys on eastern side of Naugatuck River. Deposits near Tolles were controlled by spillways at 276 m (905 ft) and 242 to 206 m (795 to 675 ft) between Naugatuck River tributaries. Deposits near Terryville are in Poland River basin and controlled by a spillway at 203 m (665 ft) into Naugatuck River tributary. Ponding in these valleys controlled by ice positions at or near interlobate angle between western side of Connecticut Valley ice lobe and western upland ice margin. (Thomaston)

emt **East Mountain Reservoir deposits**—Ice-marginal deltas ponded in west-sloping tributaries of Naugatuck River. Controlled by multiple spillways across valley divides at altitudes ranging from 221 m (725 ft) to 126 m (415 ft). Sequential ponding in these valleys controlled at interlobate angle between western side of Connecticut Valley ice lobe and western upland ice margin. (Waterbury, Naugatuck)

be **Bethany deposits**—Ice-marginal deltas ponded at multiple stages in four west-sloping tributaries of Naugatuck River. Spillways cross local divides between tributaries between 59 m (195 ft) and 194 m (635 ft). Sequential ponding in these valleys required ice-margin positions with strong north-northeast to south-southwest trend. (Naugatuck)

Farmington–Quinnipiac Basin

wbs **West Branch Salmon Brook deposits—**Ice-marginal deltas ponded at several levels in headwaters and tributaries of eastward-flowing West Branch Salmon Brook by western margin of Connecticut Valley ice lobe. Highest group of deposits is graded to spillways above 305 m (1,000 ft). A slightly younger group of deposits has spillways between 270 and 198 m (885 and 650 ft). Lowest group was impounded behind Barndoor Hills and graded to spillways at 126 m (415 ft), 117 m (385 ft), and 102 m (335 ft). (Tariffville, New Hartford)

om **Onion Mountain deposits**—Ice-marginal deltas in small north-sloping pockets east and north of Onion Mountain. Deposits are highly collapsed; only narrow fringes of noncollapsed deltaic surfaces are present. Deposits were ponded against scarp of Western Highlands by Connecticut Valley ice lobe and graded to three local spillways at 209 m (685 ft), 181 m (595 ft), and 160 m (525 ft). (Avon, Tariffville, Collinsville)

bt **Burlington deposits**—Ice-marginal deltas ponded high against edge of Western Highlands by Connecticut Valley ice lobe. Delta surfaces are successively lower to northeast and were controlled by succession of lower spillways between 288 m (945 ft) and 181 m (595 ft) across Burlington Brook–Pequabuck River drainage divide. (Collinsville)

wv **Whigville deposits**—Ice-marginal deltas ponded against collapsed slope of higher level glacial Lake Bristol deposits (lbr). Graded to two spillways at 120 m (395 ft) and 127 m (415 ft). As ice margin retreated, the lake at Whigville at 127 m (415 ft) drained; meltwater that continued to spill from Burlington lakes eroded parts of Whigville deltas. (Bristol)

qm **Quinnipiac River–Mill River divide deposits**—Ice-marginal deltas ponded in four north-draining valleys tributary to Quinnipiac River. In Tenmile River valley, deltaic surfaces are at 63 to 66 m (215 to 225 ft) and are graded to 59-m (195-ft) spillway across divide. In valley of unnamed tributary to Tenmile River, deltaic surfaces reach 62 m (205 ft) and were controlled by 56-m (185-ft) spillway. In Honeypot Brook valley, esker-fed delta has surface altitude of 63 to 69 m (205 to 215 ft) graded to 59-m (195-ft) spillway. In Broad Brook valley, there are two successive deposits; older delta has extensive collapsed ice-marginal parts south of Broad Brook Reservoir, but small noncollapsed delta surfaces at 78 m (255 ft) remain; ponding controlled by a 75-m (245-ft) spillway; younger delta north of reservoir has surface at 72 m (235 ft), controlled by 66-m (215-ft) spillway. Extensive collapsed ice-marginal parts of this deltaic sequence, including a 2-km-long (1.5-mi-long) feeder esker, block northern end of Quinnipiac Gorge and probably contributed to damming of glacial Lake Southington. (Southington, Meriden, Mount Carmel, Wallingford)

msr **Mount Sanford ridge deposits**—Ice-marginal deltas in high-level pockets in several northeast-sloping valleys that drain ridge extending north and south from Mount Sanford. Level of deposits was controlled in each valley by lowering succession of spillways that resulted from northeastward retreat of western margin of Connecticut Valley ice lobe. Spillway altitudes range from 197 m (645 ft) down to 50 m (165 ft). (Mount Carmel)

Upper Connecticut Basin

so **Somers high-level deposits**—Ice-marginal deltas in several pockets graded to local spillways that lower successively to the west from 247 m (810 ft) down to 148 m (485 ft). Ice-marginal deltaic deposits in Monson, Mass., were graded to lacustrine deposits in this series with spillway at 215 m (705 ft) east of Perkins Mountain. (Hampden)

sh **Shenipsit Lake deposits**—Ice-marginal delta at 168 m (550 ft) overlying as much as 21 m (70 ft) of lake-bottom sand and silt just south of Shenipsit Lake. Deposit was ponded by ice margin to the west and north and controlled by a spillway over older upper Connecticut River divide deposits (cd) that probably contained buried ice. North of Shenipsit Lake, deposits reach 209 m (685 ft), but have no recognizable spillway and are mostly collapsed from former levels. Deposits south of Crystal Lake straddle the eastern upper Connecticut drainage divide and were deposited from ice-margin positions east of divide by meltwater flowing southwest through local spillways. (Rockville, Ellington)

wa **Wapping deposits**—Ice-marginal deltas ponded in two pocket valleys against eastern upland before opening of glacial Lakes Middletown and Hitchcock in this area. Higher deposit was graded to two spillways at 84 m (275 ft) and 81 m (265 ft). Lower deposit has a terrace form and was ponded along ice margin and controlled by 56-m (185-ft) spillway between bedrock hills to the east. (Manchester)

vl **Vernon–Lydallville deposits**—Successive ice-marginal deltas ponded against Eastern Highlands; slightly younger and lower in altitude than northern part of upper Connecticut River divide deposits (cd). Most of these deposits are highly collapsed but some delta surfaces remain, graded to lowering successive ponding controlled by spillways from 172 m (565 ft) down to 105 m (345 ft). (Rockville)

cd **Upper Connecticut River divide deposits**—Ice-marginal deltaic deposits built into four upland lakes, impounded by eastern margin of Connecticut Valley ice lobe; lakes spilled through saddles at 154 m (505 ft), 197 to 200 m (645 to 655 ft), 187 m (615 ft), and 239 m (785 ft) across eastern part of upper Connecticut River drainage divide. Southernmost lake in this unit was previously named glacial Lake Dickinson (Langer, 1977). (Glastonbury, Rockville)

hh **Hanging Hills deposits**—Series of ice-marginal deltas in three valleys that drain northeastern slope of Hanging Hills. In Mattabesset River valley and its headwater tributaries, spillway altitudes lower northward from 114 m (375 ft) down to 53 m (175 ft). In Hatchery Brook valley, spillway altitudes range from 96 to 56 m (315 to 185 ft). In both valleys, deltaic surfaces are lower to the north, each graded to a lower spillway. In the Belcher Brook valley, deltaic surfaces are at 56 to 59 m (185 to 195 ft) and are graded to 50-m (165-ft) spillway south of Beaver Pond. (Meriden)

cg **Coginchaug River divide deposits**—Ice-marginal deltas ponded in four pocket valleys against Coginchaug River divide soon after ice margin retreated northward across it. Each delta was controlled by a spillway across the divide. Further retreat into the basin resulted in development of glacial Lake Coginchaug, represented by Durham stage deposits (lcgd) and Middletown stage deposits (lcgm), which spilled across divide at lower altitude. (Durham)

Lower Connecticut Basin

whd **West Haddam deposits**—Series of ice-marginal deltas ponded in small northeast-draining tributaries to Connecticut River. Spillways range from 172 m (565 ft) down to 102 m (335 ft) across local divides. Deposition of these small, high-level deltas in successively lower positions to the northeast was due to lobation of ice margin in lower Connecticut River valley; western side of this lobe retreated to the northeast. (Haddam)

mop **Moodus River divide–Pine Brook deposits**—Small ice-marginal deltas in northwest-sloping tributary valleys to Moodus River. Oldest deltas in each valley were graded to spillways at 108 m (355 ft), 114 m (375 ft), and 142 m (465 ft) across the southern Moodus River divide. Successive lower deltas in each valley were graded to lower spillways across smaller valley divides. Relatively extensive ice-marginal deltas in Moodus Reservoir valley and at southern end of Pine Brook valley have surface altitudes at 123 to 126 m (405 to 415 ft) and were built into small, open lake that spilled eastward over 120-m (395-ft) spillway across Moodus River divide. Unlike most deltas in this type of depositional system, these have free fronts, and lake-bottom surfaces may underlie areas now flooded by Moodus Reservoir. Deposits in north-draining Pine Brook valley, north of Babcock Pond, are series of three ice-marginal deltas graded to 117-m (385-ft) spillway that cuts through 128-m (420-ft) ice-marginal head of delta south of Babcock Pond. This lower ponding could not occur until ice margin retreated out of Moodus River valley and lake in Moodus Reservoir valley drained westward down Moodus River valley. (Moodus, Deep River)

jg **Judd Brook–Gillette Brook deposits**—Three ice-marginal deltas built in lowering succession in Judd Brook valley graded to spillways at 155 m (510 ft), 139 m (455 ft), and 126 m (415 ft). Three successive ice-marginal deltas in Gillette Brook valley east of glacial Lake Colchester graded to spillways at 160 m (525 ft), 157 m (515 ft), and 130 m (425 ft). (Colchester)

cl **Colchester deposits**—Two successive ice-marginal deltaic deposits that cross present drainage divides between upper Deep River and Lake Hayward Brook. Relationship between deposits and topography indicates that former divides were at narrows at south ends of the two deposits; divides were shifted northward after deltas were built. (Colchester)

bv **Beaver Brook deposits**—Successive ice-marginal deltaic deposits with surfaces at 29 to 47 m (95 to 155 ft) that cross present divide between Beaver Brook and Falls Brook. Initial spillway must have been across rock ridge south of Uncas Pond, but present altitude of lowest col is too low to have controlled deposits; detached ice blocks must have been present in area. Upper Cedar Pond Brook valley at northeastern end of deposit was ice free and contains fluvial feeder deposits up to 69 m (225 ft) on grade to deltaic surfaces south of Cedar Lake. (Hamburg, Old Lyme)

Thames Basin

mp **Mashapaug Pond deposits**—Ice-marginal deltas built into series of small lakes ponded against upper Quinebaug River–Shetucket River drainage divide at five successively lower spillways across divide at altitudes of 294 m (965 ft), 276 m (905 ft), 267 m (875 ft), 255 m (835 ft), and 215 m (705 ft). Deposits are particularly rusty because of abundance of sulfitic schist fragments. (Wales, Westford)

cr **Conant Brook–Roaring Brook deposits**—Ice-marginal deltas built into two groups of small lakes on eastern side of Willimantic River. Deposits in Conant Brook valley were controlled by successively lower local spillways at 219 m (720 ft), 203 m (665 ft), and 175 m (575 ft). Deposits in Roaring Brook valley were controlled by spillways across divide between two valleys at 655 ft (200 m), 635 ft (194 m), 615 ft (187 m), and 535 ft (163 m). (Stafford Springs, South Coventry)

cs **Cedar Swamp area deposits**—Successive ice-marginal deltaic deposits in Cedar Swamp valley and in north-draining tributaries to Eagleville

Brook valley, Conant Brook valley, and small tributary valleys on eastern side of Willimantic River. Earliest deposits in Cedar Swamp valley contain extensive ice-channel fillings and were controlled by 169-m (555-ft) spillway; these deposits blocked the valley. Successive deposits north of Cedar Swamp were controlled by 178-m (585-ft) spillway. (South Coventry)

fv **Fitchville deposits**—Ice-marginal deltas in seven north-draining tributaries to Yantic River; single deltas in each valley were controlled by spillways across each local divide. (Fitchville, Norwich)

gl **Gardner Lake lacustrine deposits**—Successive ice-marginal deltaic deposits in north-draining Deep River valley and in Gardner Brook valley where deposits bury former divide. Deltaic deposits north of Gardner Lake have surface altitudes of 123 to 126 m (405 to 415 ft) and were built into small lake controlled by 120-m (395-ft) spillway. Deltas in Deep River valley, including one near-ice-marginal delta, have surface altitudes of 130 m (425 ft) and were built into small lake controlled by 127-m (415-ft) spillway across divide at The Wales. Slightly lower deltas on northern side of Deep River Reservoir are extensively collapsed and were controlled by 117-m (385-ft) spillway across deposits occupying divide area north of Gardner Lake. (Fitchville, Colchester)

led **Ledyard deposits**—Ice-marginal deltas in three north-draining tributaries to Thames River. Deltas with surfaces at 50 to 60 m (165 to 185 ft) in Billings Avery Brook valley contain numerous well-developed kettles and were controlled by two spillways that cross Ledyard moraine (hlm) and head of Groton deposits (gr) at 53 m (175 ft) and 47 m (155 ft). Deltas in Joe Clark Brook and Shewville Brook valleys to the north were controlled by three local spillways at 53 m (175 ft), 47 m (155 ft), and 38 m (125 ft). (Uncasville)

Quinebaug and Southeast Coastal Basin

nth **North Thompson deposits**—Ice-marginal deltas with esker feeder deposits in two north-sloping tributary valleys. Eastern valley is tributary to Fivemile River valley and has delta surfaces at 157 m (515 ft), controlled by 151-m (495-ft) spillway across divide; western valley is tributary to French River valley and has two successive deltas at 154 m (505 ft) and 151 m (495 ft), controlled by 145-m (475-ft) spillway. (Webster, Oxford)

epg **East Putnam–Glocester deposits**—Successive ice-marginal deltas in northwest-sloping Cady Brook and Mary Brown Brook valleys. Series of ponds began in Glocester, R.I., controlled by two spillways across Quinebaug River divide; first spillway is in Rhode Island, second spillway at 178 m (585 ft) is in Connecticut. Later spillways for ponding in Connecticut descend from 142 m (465 ft) to 130 m (425 ft). (Thompson)

wb **White Brook deposits**—Four successive ice-marginal deltas in White Brook valley; deposits are extensively collapsed. Series of ponds controlled by spillways at 93 m (305 ft), 90 m (295 ft), and 84 m (275 ft) across local divide. (Danielson)

ms **Mashentuck Brook deposits**—Ice-marginal deltas in north-draining Mashentuck Brook valley. Ponding controlled by two spillways at 130 m (425 ft) and 123 m (405 ft). (East Killingly)

wr **Wauregan deposits**—Three successive ice-marginal deltaic and deltaic-fluvial deposits in north-sloping tributary valley to Quinebaug River. Ponding was controlled by two 69-m (225-ft) spillways across southern divide. Highly collapsed deposits that reach 87 m (285 ft) in northern part of unit are chiefly fluvial. (Danielson, Plainfield)

mb **Mill Brook deposits**—Four successive ice-marginal deltas in north-draining Mill Brook valley. Sediment was deposited around numerous stagnant ice blocks. Levels of ponding were controlled by spillways at 53 m (175 ft) and 50 m (165 ft) across divide at southern end of unit. (Plainfield, Jewett City)

bd **Broad Brook deposits**—Two successive ice-marginal deltas in lower part of north-draining Broad Brook valley. Level of ponding was controlled by 50-m (165-ft) spillway across local divide. (Jewett City)

bm **Bay Mountain deposits**—Successive ice-marginal deltas related to nine ice-margin positions in three relatively steep, north-sloping valleys in Bay Mountain upland area. Eastern valley occupied by Billings Brook contains most extensive deposits and is tributary to Pachaug River basin. Earliest and highest levels of ponding in this system controlled by spillway at 99 m (325 ft) across Quinebaug River basin divide; some later levels controlled by spillways across local divides, but downstream from these, path of meltwater flow was across basin divide through 69-m (225-ft) spillway. (Jewett City, Old Mystic)

Deposits of Related Series of Sediment-Dammed Ponds—SP Depositional System

Paleogeographic setting: Many valleys in Connecticut sloped to the south away from the ice margin. In narrower sections of these south-sloping valleys, series of small lakes developed sequentially as a result of northward ice retreat. Each pond was dammed behind (to the north of) the valley-blocking body of sediment that filled the next previous pond. In steeper sections of these valleys, meltwater streams fed a small lake farther down the valley. Spillways for each small lake were over sediment dams; these spillways commonly no longer exist because most of the sediment was removed by distal meltwater and ancestral streams in each valley. The process of degradation and entrenchment of ice-marginal deposits in these narrow south-sloping valleys was aided by a lowering base level (glacial Lake Connecticut) in Long Island Sound.

Deposits: Predominantly deltaic and fluvial sediments; lake-bottom sediments occur locally beneath deltaic sediments but are not exposed at the surface. Ice-marginal deltaic, fluviodeltaic, and deltaic-fluvial morphosequences are present; near-ice-marginal fluviodeltaic deposits occur rarely.

Stratigraphic arrangement of deposits: Deltas in this depositional system generally do not have free fronts; deltaic sediments commonly filled small ponds and are contiguous with the ice-marginal heads of previous morphosequences in the valley. Lacustrine sediments consist mostly of bottomset beds and are present only beneath coarser deltaic sediments. Deltaic surfaces and topset-foreset contacts within a map unit commonly rise up valley at a rate greater than the slope of postglacial tilt. Because of thick fluvial-topset aggradation in each deltaic sediment dam, levels of ponding were successively higher up valley. Fluvial sediments within fluviodeltaic morphosequences occur in steeper sections of some valleys where the till and (or) bedrock floor is at shallow depths. Deltaic-fluvial morphosequences occur in sections of gentler gradient in some valleys; in these places, fluvial beds overlie deltaic beds and may overlap from one morphosequence to other ice-marginal deltas farther down the valley. In many valleys, only remnant deposits of the SP depositional system remain, especially in longer river valleys of the Eastern and Western Highlands such as the Thames, lower Connecticut, Naugatuck, and Housatonic.

Map units: Shown in brown colors on the map. Unit descriptions (below) include information on location of the valley or valleys in which small glacial lakes existed; types of morphosequences and their distribution within the valley or valleys; ranges of delta-surface altitudes and fluvial-surface gradients across the groups of morphosequences within map units. Most unit descriptions include reference to the cause of initial ponding, and the circumstances under which the particular lake series ended and a new one began.

sp **Uncorrelated deposits of sediment-dammed ponds**

Housatonic–Southwest Coastal Basin

sy **Salisbury deposits**—Ice-marginal fluvial deposits reach as high as 230 m (755 ft) in northern part of unit; grade southward to ponded sediments in Salisbury village area. Near-ice-marginal fluviodeltaic deposits built from the west overlie lake-bottom deposits at Lakeville. Ponding probably was in overdeepened bedrock basin behind Salmon Creek narrows at southern end of unit. (Bashbish Falls, Sharon)

hrn **Housatonic River deposits from New Milford to West Cornwall**—Sequential ice-marginal deltaic and fluviodeltaic deposits along Housatonic River between New Milford village and West Cornwall; fluvial feeder deposits in some steeper tributary valleys to Housatonic River. Numerous ice-margin positions identified, although deposits in many places are only discontinuous remnants due to extensive distal-meltwater and postglacial erosion. Deltaic surfaces are at 81 m (265 ft) at southern end of unit and 184 m (605 ft) at northern end; shingled-profile breaks identified in many places in association with ice-margin positions. Ponding in this part of valley began immediately following drainage of last stage of glacial Lake Danbury; initial impoundment was behind glacial Lake Danbury deposits. (New Milford, Kent, Dover Plains, Ellsworth, Cornwall, South Canaan)

hrs **Housatonic River deposits from Stratford to Shepaug River**—Discontinuous remnants of sequential ice-marginal deltaic deposits along Housatonic River between Stratford and mouth of Shepaug River in Southbury. Surface altitudes at 17 m (55 ft) at southern end of unit rise with minor identifiable shingled-profile breaks to 78 m (255 ft) at northern end of unit. Deposits were considerably incised by later meltwater and postglacial erosion so that reconstruction of surface gradients, ice-margin positions, and delta surfaces is precluded. Initial ponding in valley began directly following deposition of Stratford–Southport deposits (lcss) into glacial Lake Connecticut at mouth of valley. (Milford, Ansonia, Long Hill, Southbury, Newtown) sr **Shepaug River deposits**—Largely discontinuous remnants of sequential ice-marginal deltaic deposits along lower Shepaug River. Probably related to at least three ice-marginal positions with levels of ponding between 72 m (235 ft) and 84 m (275 ft). Initial ponding behind last deposits of Housatonic River (hrs) in Housatonic River valley. (Roxbury, Newtown)

eb **Eightmile Brook deposits**—Several ice-marginal deltaic and fluviodeltaic deposits in Eightmile Brook and Little River valleys; northern sequences have extensive fluvial sections. Initial ponding in southern part of unit was behind till and (or) bedrock spillway on side of valley; successive sequences controlled by sediment-dammed ponding. Meltwater deposits also spilled southeast into upper part of Little River valley. (Southbury)

pf **Pequonnock River–Farmill River deposits**—Ice-marginal and near-ice-marginal fluviodeltaic and deltaic-fluvial deposits in upper reaches of Mill, Pequonnock, and Farmill River valleys and their tributaries. Ponding in many places was in overdeepened basins behind bedrock narrows; in other places, such as at Stepney in Pequonnock River valley, sequential ponding was behind successive ice-contact heads in valley. Meltwater construction of this system in upper reaches of these valleys closely followed construction of fluviodeltaic Stratford-Southport deposits (lcss) that graded to glacial Lake Connecticut in lower reaches of valleys. (Botsford, Long Hill, Bridgeport)

sas **Saugatuck River–Aspetuck River deposits**—Near-ice-marginal fluviodeltaic and deltaic-fluvial deposits in upper Saugatuck River valley and tributary Aspetuck River valley. Initial ponding was behind head of Stamford–Norwalk–Westport deposits (lcsnw) at Westport; subsequent ponding was mostly in wider basins behind bedrock narrows with only minor sediment damming. Deposits in West Branch Saugatuck River valley included in this unit are largely fluvial, but southern part may have been tributary feeder to deltaic deposits at Saugatuck River valley confluence. (Botsford, Norwalk North, Westport, Norwalk South, Sherwood Point)

nu **Upper Norwalk River deposits**—Sequential near-ice-marginal fluviodeltaic deposits in upper Norwalk River and Silvermine River valleys. Fluvial deposits in narrower and steeper sections of valleys grade to deltaic deposits ponded in overdeepened basins behind bedrock narrows; successive northward construction of this depositional system indicated by distribution of textures within deposits rather than shingled-surface profiles. (Bethel, Norwalk North)

rn **Rippowam River–Noroton River deposits**—Sequential near-ice-marginal fluviodeltaic deposits in Rippowam and Noroton River valleys. Fluvial deposits in narrower and steeper sections of valleys grade to deltaic and lacustrine deposits in deeper basins near Interstate Route 95. Successive northward construction within this unit indicated by textural distribution rather than shingled-surface profiles. (Pound Ridge, Stamford)

Naugatuck Basin

nrt **Naugatuck River deposits from Thomaston to Litchfield**—Near-ice-marginal fluviodeltaic deposits in southern half of unit. Ponded initially behind head of Naugatuck River deposits from Naugatuck to Reynolds Bridge (nrn) at about 122 m (400 ft). Ice-marginal fluvial deposits of very coarse gravel at northern end of unit reach 152 m (500 ft) and grade southward to deltaic deposits at 133 m (435 ft) just north of Thomaston Dam. (Thomaston)

nrn **Naugatuck River deposits from Naugatuck to Reynolds Bridge**—Sequential ice-marginal deltaic and fluviodeltaic deposits and near-ice-marginal fluviodeltaic deposits in Naugatuck River from Naugatuck northward to Reynolds Bridge. Earliest deposits in this unit are fluviodeltaic deposits built in two tributary valleys to Naugatuck River (Beacon Hill Brook from east and Long Meadow Pond Brook from west). Early levels of ponding at 96 m (315 ft) must have been controlled by drift and (or) ice in narrow bedrock gorge of Naugatuck River south of unit. Inset against higher ponded deposits in main valley at Naugatuck are deltaic deposits at 81 m (265 ft). Delta surfaces rise northward through this section of valley to 120 m (395 ft), a gradient of about 1.5 m/km (8 ft/mi) (0.8 m/km (4 ft/mi) when postglacial tilt is taken into account). (Thomaston, Waterbury, Naugatuck)

mdr **Mad River deposits**—Ice-marginal and near-ice-marginal fluviodeltaic deposits in Mad River valley built from ice both in upper Mad River and along Naugatuck River. Initial spillway at

136 m (445 ft) across East Mountain Reservoir deposits (emt) in Hopeville Brook valley. Later spillway across earlier head at Mill Plain controlled ponding levels in northern part of unit at about 137 m (450 ft). (Waterbury, Southington)

nrd **Naugatuck River deposits from Derby to Beacon Falls**—Ice-marginal and near-ice-marginal fluviodeltaic deposits in Naugatuck River valley from confluence with Housatonic River northward to Beacon Falls. Fluvial feeder deposits reach 72 m (235 ft) in short tributary valleys and grade to deltaic deposits in main valley. Levels of ponding recorded in deltas are about 30 m (100 ft) at southern end of unit and about 61 m (200 ft) at Beacon Falls, a rise of approximately 1.9 km/km (10 ft/mi) (1.2 m/km (6 ft/mi) when postglacial tilt is taken into account). (Naugatuck, Ansonia)

Farmington–Quinnipiac Basin

bk **Barkhamsted Reservoir deposits**—Highly collapsed deposits south of Saville Dam. Maximum altitudes reach 149 m (490 ft) and probably are deltaic, ponded behind 152-m (500-ft) upper Farmington River deposits (fu) south of Puddle Town. North of Saville Dam, only scattered probable deltaic surfaces are visible above reservoir water level; altitudes are at 172 m (565 ft) and reach 178 m (585 ft) at ice-margin position just north of Barkhamsted town line. North of this position, deposits with surfaces at 166 m (545 ft) rise northward to 181 m (595 ft). North of reservoir, presumably fluvial deposits with 198-m (650-ft) surfaces continue on steep gradient into Massachusetts. (New Hartford)

ss **Southwick–Suffield deposits**—Successive ice-marginal deltaic deposits with delta surfaces at 78 to 81 m (255 to 265 ft); initial spillways were over heads of deposits of glacial Lake Tariffville (lt). Deposits continue northward into Massachusetts where massive ice-marginal deltas surrounding Congamond Lakes have surfaces that reach 84 m (275 ft). (Tariffville, Southwick, West Springfield)

fu **Upper Farmington River deposits**—Ice-marginal deltaic deposits in upper Farmington River valley that were highly dissected by later meltwater and by Farmington River. Remaining terraces are between Collinsville and New Hartford. The 152-m (500-ft) delta south of Puddle Town extends across valley. From New Hartford northwest along West Branch Farmington River, surfaces rise on rather steep gradient and are predominantly fluvial with possible deltaic portions in Hartland. (Collinsville, New Hartford, Winsted)

sl **Simsbury deposits**—Successive ice-marginal deltaic deposits built in part against stagnant ice along narrow valley of present Farmington River, resulting in collapsed distal deltaic sands southwest of Simsbury village. Older deltaic surfaces reach 102 m (335 ft), younger surfaces are at 90 m (295 ft). As much as 50 m (165 ft) of lake-bottom silt and clay underlies Farmington River alluvium. Levels of ponding were controlled by massive ice-contact head of glacial Lake Farmington deposits (lf) that blocked valley to the south at maximum altitude of 84 m (275 ft). (Tariffville, Avon)

ws **West Simsbury deposits**—Ice-marginal deltaic and fluviodeltaic deposits. Westernmost deposits in Avon quadrangle at 93 to 111 m (305 to 365 ft) are mostly fluvial. Slightly younger deposits to the east are successively ponded deltaic sediments rising 87 m (285 ft) in northern Avon to 102 m (335 ft) northward into Simsbury. In Tariffville quadrangle, consists of highly collapsed remnants of ice-contact head and narrow fluvial terrace that reaches 110 m (360 ft) at north end. (Avon, Tariffville)

un **Unionville deposits**—Fluvial deposits in Bristol. Distal sands overlie gravelly ice-marginal head of Southington–Bristol deposits (sb) at East Bristol; this fluvial morphosequence rises northward to ice-contact head in southeastern corner of Burlington near Lake Como with nearly 3 km (2 mi) of esker feeders. Behind this head, consists of deltaic deposits near Lake Garda; these have an ice-marginal head at 99 to 105 m (325 to 345 ft) along southwestern side of Farmington River. On northeastern side of river, scattered remnants of ice-contact deposits remain near Unionville and range from 96 to 114 m (315 to 375 ft); these are probably deltaic, built in and around stagnant ice while meltwater still escaped southwestward via Unionville Brook–Cherry Brook valley. North of river, six or seven eskers descend down southwestern slope towards river and are some of the few examples of southward-sloping eskers in Connecticut; may have been built in subglacial tunnels or collapsed down onto southward-sloping till surface from high in the ice. (Bristol, Collinsville, Avon) mu **Upper Mill River deposits**—Predominantly successive ice-marginal deltaic deposits. Surface gradient over whole unit is relatively gentle, rising from 35 to 59 m (115 to 195 ft) with head at West Cheshire reaching 67 m (220 ft). Deposits were initially ponded behind bedrock narrows at Mount Carmel and head of New Haven deposits (lcnh). Ponding continued as ice retreated northeastward because of blockage by at least two ice-contact heads, first near Hamden-Cheshire town line and second at West Cheshire. (Mount Carmel)

sb **Southington–Bristol deposits**—Successive ice-marginal deltas, several of which have esker feeders in Eightmile River and Patton Brook, tributary valleys to Quinnipiac River. Delta surfaces rise from 69 m (225 ft) to 78 m (255 ft) in the Eightmile River valley and to 75 m (245 ft) in Patton Brook valley. Deltas were ponded behind heads of last deltas built into open glacial Lake Southington. (Bristol, New Britain, Southington, Meriden)

fm **Upper Farm River–Muddy River deposits**—Two successive fluviodeltaic morphosequences in upper Farm River valley: the first was ponded behind the 34-m (110-ft) ice-marginal head of East Haven deposits (lcenh) and surfaces rise from 29 m (95 ft) to head at 53 m (175 ft); ponding occurred behind this head also, and surfaces rise from 50 m (165 ft) up to 61 m (200 ft) northeast of Northford. Deposits in Muddy River valley consist of probably three successive fluviodeltaic deposits; fluvial sediments predominate and are graded to less extensive deltaic sediments north of valley narrows. (Branford, Wallingford)

wu **Upper Wepawaug River deposits**—Mostly fluvial sediments with ponded sediments occurring at southern end of unit around Wepawaug Reservoir. Ponding was caused by presence of ice to the southeast. Western margin of Connecticut Valley ice lobe provided dam as well as feeding meltwater streams for this unit. (Ansonia)

Upper Connecticut Basin

db **Dividend Brook deposits**—Successive ice-marginal deltas with surface altitudes of 47 to 50 m (155 to 165 ft) in southern part of unit and 53 to 56 m (175 to 185 ft) in northern part. Topset-foreset contact in Mustard Bowl pit just north of Cromwell-Rocky Hill town line is estimated at 44 to 45 m (146 to 149 ft). Deltas built into a small glacial lake that was temporarily ponded to a slightly higher level than glacial Lake Middletown behind ice-marginal deltas of Cromwell delta deposits (lmc) and before that sector of the ice margin retreated north of Cedar Mountain; water level was controlled by Dividend Brook spillway across delta surface of Cromwell delta deposits (lmc) and lowered from about 46 m (150 ft) to 40 m (129 ft) during the life of the Dividend Brook lake. (Hartford South, Glastonbury)

Lower Connecticut Basin

hu **Upper Hammonasset River deposits**—Series of probably four ice-marginal and near-ice-marginal fluviodeltaic morphosequences in upper Hammonasset River valley. Morphosequence deposition was controlled by topographic basins along valley separated by bedrock narrows formerly occupied by sediment dams. Surface altitudes over unit as a whole range from 84 m (275 ft) northward to 98 m (320 ft); ice-margin positions are recorded in two places by slight increases in surface altitudes and coarser materials; behind each position, sediments are finer and altitudes slightly lower. (Durham, Haddam)

hr **Hammonasset River deposits**—Ice-marginal deltas initially ponded behind head of Hammonasset River–Menunketesuck River deposits (lchm) in lower Hammonasset River valley. Near-ice-marginal fluviodeltaic sequence in northern part of unit; fluvial deposits reach 46 m (150 ft) in Chatfield Hollow valley and grade southward to deltaic surfaces at 26 m (85 ft). (Clinton)

qn **Quonnipaug Lake–East River–West River deposits**—Series of near-ice-marginal fluviodeltaic deposits, initially ponded behind heads of East River–West River deposits (lcew) in lower parts of East and West Rivers. To the north, ponding occurred in topographic basins behind sediment-filled bedrock narrows. Predominantly fluvial sediments occur in steep, narrow sections of East and West Rivers. (Durham, Guilford)

bfr **Upper Branford River deposits**—Series of ice-marginal fluviodeltaic deposits. Ponding occurred mostly in topographic basins behind sediment-filled bedrock narrows. (Branford, Guilford)

wo **Westbrook–Old Saybrook deposits**—Ice-marginal deltaic and fluviodeltaic deposits in upper Patchogue and Oyster River valleys and their tributaries. Ponded initially behind Westbrook–Old Saybrook–Old Lyme deposits (lcwoo). Fluvial feeders reach 18 to 21 m (60 to 70 ft) altitude; deltaic surfaces at 11 to 14 m (35 to 45 ft). (Essex)

lce **Lower Connecticut River (Essex–Hamburg–Deep River) deposits**—Predominantly ice-marginal deltaic deposits of which only remnants remain due to downcutting by postglacial Connecticut River. Delta surfaces are at 11 m (35 ft) in southern part of unit and 14 to 17 m (45 to 55 ft) in northern part. Near-ice-marginal fluviodeltaic sequence in Joshua Creek valley in Lyme has fluvial surfaces that reach 21 m (70 ft) and grade to 14-m (45-ft) delta surface along Connecticut River. Initial ponding in lower Connecticut River valley was behind Westbrook–Old Saybrook–Old Lyme deposits (lcwoo) at mouth of river. (Hamburg, Old Lyme, Essex, Deep River)

lcc **Lower Connecticut River (Chester–Hadlyme) deposits**—Near-ice-marginal fluviodeltaic deposits in Pattaconk Brook–Chester Creek valley in Chester, in Roaring Brook–Whalebone Creek valley in Hadlyme, and along Connecticut River. Fluvial deposits with surfaces as high as 30 m (100 ft) in Chester and 38 m (125 ft) in Roaring Brook valley in Hadlyme grade eastward and westward, respectively, to deltaic surfaces on either side of Connecticut River valley at 17 to 20 m (55 to 65 ft). Ponding to this level in Connecticut River valley was due to blockage by sediment dams to the south formed by Westbrook–Old Saybrook–Old Lyme deposits (lcwoo) and Essex–Hamburg–Deep River deposits (lce). Also included are small, near-ice-marginal fluviodeltaic morphosequences in upper sections of Hemlock Valley Brook, Hungerford Brook, and Roaring Brook valleys in Hadlyme. Each deposit occupies topographic basin in the valley and is graded to local tributary-valley spillway at its southern end; each spillway is somewhat higher than present stream gradient through valley narrows; present stream courses were probably blocked by till (or possibly buried ice) at time of meltwater deposition. (Deep River)

lct **Lower Connecticut River (Tylerville–Portland) deposits**—Predominantly ice-marginal deltaic deposits; probably more successive morphosequences exist than are indicated by ice-margin position symbols on map. From Tylerville to north of Haddam-Middletown town line, deltaic surfaces rise from 23 m (75 ft) to 34 m (110 ft) (a gradient of about 1.2 m/km (6 ft/mi), most of which can be accounted for by postglacial uplift); fluvial topset beds are generally thin (less than 3 m (10 ft) thick). Deltaic surfaces at and just south of Maromas are at 41 to 44 m (135 to 145 ft) and rise to 53 m (175 ft) at northernmost ice-marginal head; fluvial topset gradients on these deposits are steeper (about 3 m/km (10 ft/mi)) and topset beds are as much as 8 m (25 ft) thick. Deposition of greater thicknesses of topset beds in northern ice-marginal delta series of this unit is perhaps result of slower retreat in this area, where regional margin of ice sheet was mainly retreating down opposing northwest-facing slopes. Northernmost deposits, at Jobs Pond north of Connecticut River, are as much as 76 m (250 ft) thick and block former channel of Connecticut River; because these deposits were not trenched by later erosion, top of next-to-the-last delta preserves spillway channel at 47 m (155 ft) (same altitude as topset-foreset contact in last delta). This illustrates self-damming effect characteristic of SP depositional system; this type of spillway was not commonly preserved. (Middle Haddam, Deep River, Haddam)

ol **Old Lyme deposits**—Ice-marginal deltaic deposits in upper Black Hall River valley with surfaces at 14 to 17 m (45 to 55 ft). Near-ice-marginal fluviodeltaic deposits in northern tributary valleys to Lieutenant River valley; fluvial deposits reach 21 m (70 ft) north of Rogers Lake and grade to deltaic surfaces south of lake at 14 to 17 m (45 to 55 ft). Deposits ponded behind heads of Westbrook–Old Saybrook–Old Lyme deposits (lcwoo) in Black Hall and Lieutenant River valleys. (Old Lyme)

eg **Eightmile River deposits**—Successive ice-marginal deltas in Eightmile River valley with surfaces rising to the north from 17 m (55 ft) to 35 m (115 ft). Initially ponded behind Essex–Hamburg–Deep River deposits (lce) in Hamburg Cove; near-ice-marginal fluviodeltaic deposits north of State Route 82 have deltaic surfaces at 35 to 38 m (115 to 125 ft) near North Plain and fluvial surfaces rising to 53 m (175 ft) at northern end of unit. (Hamburg)

sa **Salmon River deposits**—Predominantly near-ice-marginal fluviodeltaic deposits in lower Salmon River valley; one ice-margin position recorded at northeastern end of unit. Surface altitudes range from 29 m (95 ft) in the south to 59 m (195 ft) at northern end of unit. Because northeast-southwest trend of Salmon River valley was parallel to retreating ice margin, southern part at least probably was uncovered all at once instead of sequentially northward as were valleys that trended perpendicular to ice margin. Initial ponding of Salmon River valley was behind part of Tylerville–Portland deposits (lct), which provided a sediment dam; as ice margin retreated northwestward out of Salmon River valley, meltwater flowed down northwest tributary valleys of Pine, Elbow, and Wopowog Brooks. Fluvial sediments were deposited in steeper sections and these grade downstream to deltaic sediments in main valley; these deltas must have filled and blocked sections of valley, resulting in higher levels of ponding up Salmon River valley, as is indicated by 30-m (100-ft) rise in altitudes of deltaic surfaces. (Moodus, Deep River, Middle Haddam)

bl **Lower Blackledge River deposits**—Series of ice-marginal deltaic deposits in main valley and near-ice-marginal fluviodeltaic deposits in which fluvial sediments in tributary valleys grade to deltas in main valley. Initial ponding was behind Salmon River deposits (sa) in Salmon River valley; levels of ponding increased up valley. Delta surfaces rise from 64 m (210 ft) at southern end of unit (ponded behind 67-m (220-ft) head of unit sa) to 99 m (325 ft) at northern end. (Marlborough, Moodus)

Thames Basin

mr **Middle River deposits**—Ice-marginal deltaic and fluviodeltaic deposits ponded behind blockage in bedrock narrows at Stafford Springs village. Deltaic surfaces are at 166 m (545 ft) at southern end of unit; fluvial feeder deposits, including eskers, reach 184 m (605 ft) in tributary valley of Patten Brook and at ice-marginal head just south of State Line Pond. Fluviodeltaic deposits extending into Massachusetts were ponded behind 184-m (605-ft) head. (Monson, Stafford Springs)

bg **Bigelow Brook deposits**—Ice-marginal deltaic deposits at southern end of unit at 171 m (560 ft); blocked initially by uncorrelated ice-dammed pond deposits of unit ip. Terrace remnants at 188 m (615 ft) rise northward to ice-marginal position at 203 m (665 ft) in central part of unit; these deposits probably are mostly fluvial. Ice-marginal deltaic deposits with long feeder esker at northern end of unit extend across divide into Quinebaug River basin. (Southbridge, Westford, Eastford)

sk **Skungamaug River deposits**—Ice-marginal fluviodeltaic deposits built as two successive sequences in southern half of valley graded to ponded level of about 148 m (485 ft); lake-bottom deposits at southern end with surface level of 137 to 140 m (450 to 460 ft); valley probably was dammed by meltwater deposits (u) in valley narrows at southern end. Northern half of unit was ponded behind moraine deposits (m) 2 km (1.6 mi) south of Tolland village. Deltaic surfaces are at 166 m (545 ft) in this section of valley; fluvial deposits reach 175 m (575 ft) at northern end. Sandy surfaces at 155 to 158 m (510 to 520 ft) just north of moraine probably are lake-bottom deposits. (Stafford Springs, South Coventry)

hp **Hop River deposits**—Succession of at least six ice-marginal deltaic morphosequences along Hop River valley, although only two ice positions are clearly evident from surface morphology. Surface gradients rise from 90 m (295 ft) at eastern end of valley where deposits merge with those in Willimantic River valley, to 123 m (405 ft) to the west; this is an overall gradient of about 0.8 m/km (4 ft/mi), some of which is result of postglacial rebound. (Columbia, Marlborough, Rockville)

mf **Mount Hope River–Fenton River deposits**—Successive ice-marginal fluviodeltaic deposits along two river valleys. Coarse gravelly terrace deposits and still coarser esker segments may be mostly fluvial with only minor deltaic sections; these deposits almost certainly are composite, but ice-contact heads are not clearly shown. At southern end of unit, where two valleys join in Mansfield Hollow Lake area, extensive deltaic deposit at 81 m (265 ft) was dammed behind northernmost ice-marginal head of Willimantic River–Upper Shetucket River deposits (wil). (Spring Hill)

na **Natchaug River deposits**—Successive ice-marginal fluviodeltaic deposits in Natchaug River valley; ice-margin positions not clearly shown. Deltaic surfaces rise from 111 m (365 ft) at southern end of unit to more than 183 m (600 ft) at northern end. This relatively steep gradient indicates much fluvial deposition. (Eastford, Hampton)

wru **Upper Willimantic River deposits**—Series of ice-marginal fluviodeltaic deposits in section of upper Willimantic River valley that is narrow with a steep gradient. Ponding was initially behind head of Willimantic River deposits (wrl) at about 96 m (315 ft); deltaic surfaces are as high as 126 m (415 ft) in northern part of unit. (Stafford Springs, South Coventry)

wrl **Willimantic River deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits in narrower section of Willimantic River valley behind head of Willimantic River–upper Shetucket River deposits (wil). Related to at least two ice-margin positions. Levels of ponding are slightly higher than in unit wil at 84 to 90 m (275 to 295 ft); fluvial feeder deposits reach 93 m (305 ft). (South Coventry, Columbia)

wil **Willimantic River–upper Shetucket River deposits**—Predominantly successive ice-marginal deltaic and fluviodeltaic deposits in relatively wide, shallow-gradient section of Shetucket River valley. Deposits are related to at least 12 ice-margin positions. Initial ponding was behind lower Shetucket River deposits (str) and remnant ice blocks in narrows of Shetucket River at southern end of unit. Levels of successive ponding were at 72 to 78 m (235 to 255 ft); fluvial feeder deposits reach as high as 91 m (300 ft) in lower Natchaug River valley and Potash Brook valley. Ice-margin-parallel ridge at head of first ice-marginal delta in this series is at 84 m (275 ft), 15 m (50 ft) above the 72-m (235-ft) delta surface in front of it; this feature possibly may be a morainic ridge, but no gravel pits exist in which to observe internal structure. Relatively extensive meltwater-terrace unit inset into deltaic surfaces at Windham records 5- to 6-m (15- to 20-ft) drop in ponding level while meltwater still flowed down Potash Brook valley. (Willimantic, Scotland, Spring Hill)

lru **Upper Little River deposits**—Ice-marginal deltaic and deltaic-fluvial deposits built at three southern ice positions; northernmost sequence is near-ice-marginal fluviodeltaic deposit. Ponding in southern part was at 108 to 111 m (355 to 365 ft), initially behind head of lower Little River deposits (lrl); fluvial deposits in third sequence reach as high as 142 m (465 ft); northern sequence ponded behind this head at 130 to 133 m (425 to 435 ft) has fluvial feeder deposits as high as 163 m (535 ft) at northern end. (Hampton, Scotland)

lrl **Lower Little River deposits**—Ice-marginal deltaic-fluvial morphosequences in southern half of Little River valley, related to at least three ice-margin positions. Initial ponding was at a level of 44 to 47 m (145 to 155 ft) behind lower Shetucket River deposits (str) at Versailles; surface gradients rise steeply northward to 111 m (365 ft) at northern end of unit. Levels of ponding increase about 15 m (50 ft) between sequences because of relatively thick fluvial aggradation at each head. (Scotland, Norwich)

mbb **Merrick Brook–Beaver Brook–Ballymahack Brook deposits**—Near-ice-marginal fluviodeltaic deposits in three valleys with long fluvial sections of very coarse gravel; fluvial deposits are as high as 137 m (450 ft) and grade southward with 7.8 to 9.7 m/km (40 to 50 ft/mi) gradients to deltaic deposits at 69 m (225 ft). Deltaic-fluvial deposits initially ponded against local divides at heads of Beaver and Merrick Brooks, near the long esker feeder; fluvial aggradation continued across those divides. (Spring Hill, Scotland, Willimantic, Hampton)

str **Lower Shetucket River deposits**—Successive ice-marginal deltaic deposits in main Shetucket River valley; ice-marginal and near-ice-marginal fluvial deposits in tributary valleys grade to deltas in Shetucket River and Beaver Brook valleys. Initial ponding was behind Yantic River deposits (yr) in Thames River valley; delta altitudes increase from 29 m (95 ft) at Taftville to 59 m (195 ft) at northern end of unit. (Norwich, Fitchville, Willimantic)

yr **Yantic River deposits**—Ice-marginal deltas and fluviodeltaic deposits; near-ice-marginal fluvial deposits in tributary valleys are on grade to deltas in main Yantic River valley. Initial ponding was behind deltas of glacial Lake Uncasville (lu) and Trading Cove Brook (tc) in Thames River valley; delta altitudes rise from 29 m (95 ft) at confluence of Yantic and Thames Rivers to 53 m (175 ft) near village of Fitchville. (Norwich, Fitchville)

tc **Trading Cove Brook deposits**—Near-ice-marginal fluviodeltaic and ice-marginal deltaic deposits; initially ponded behind deposits of glacial Lake Uncasville (lu) in Thames River valley at Trading Cove. Fluvial deposits in tributary valleys of Ford, Gardner, and Great Plain Brooks are as high as 35 m (115 ft) and grade to delta surfaces in lower Trading Cove Brook valley at 23 to 26 m (75 to 85 ft); northern part of unit consists of fluvial deposits as high as 72 m (235 ft) at northern end that slope steeply to delta surface at 32 m (105 ft) behind earlier ice-marginal deltas. (Fitchville, Norwich, Uncasville)

ox **Oxoboxo Brook deposits**—Ice-marginal fluviodeltaic deposits in southeastern end of unit and successive ice-marginal deltas in northwestern part of Oxoboxo Brook valley. Proximal fluvial deposits at southern end of Oxoboxo Lake, probably built at Oxoboxo moraine (mom) position, are at 133 m (435 ft) and slope down-valley with a surface gradient greater than 9.7 m/km (50 ft/mi) to deltaic deposits at 84 m (275 ft) behind bedrock narrows at southern end of Wheeler Pond and head of glacial Lake Uncasville deposits (lu). Ice-marginal deltas north of Oxoboxo Lake were ponded behind the 133-m (435-ft) fluvial deposits and have surface altitudes of 127 to 130 m (415 to 425 ft). (Fitchville, Montville)

el **East Lyme deposits**—Ice-marginal and near-ice-marginal fluviodeltaic deposits in Fourmile River, Pataguanset River, and Stony Brook valleys; ponding in each valley was initially behind Niantic deposits (lcn). Delta surface altitudes increase from 20 m (65 ft) to 41 m (135 ft) in Fourmile River valley and from 14 m (45 ft) to 32 m (105 ft) in Pataguanset River valley and its tributaries, Latimer Brook and Lakes Pond Brook; youngest deposits head at Ledyard moraine (hlm). (Old Lyme, Niantic, Montville)

Quinebaug and Southeast Coastal Basin

fr **Fivemile River deposits**—Four successive ice-marginal deltas with surfaces at 123 to 126 m (405 to 415 ft) in valley south of Quaddick Reservoir; probably controlled by 123-m (405-ft) spillway on valley wall just downstream, at a time when narrow valley was blocked. Ice-marginal deltaic and fluviodeltaic deposits in northern half of unit have surface altitudes that rise at steeper gradient, from 133 m (435 ft) to 160 m (525 ft). Behind last ice-margin position in Connecticut, successive ice-marginal deltas through Lake Chaubunagungamaug basin and extending into Massachusetts are at 154 to 160 m (505 to 525 ft) and were graded to a spillway preserved at 145 m (475 ft) across earlier head. (Oxford, Thompson)

wt **West Thompson deposits**—Remnants of successive ice-marginal fluviodeltaic deposits in Quinebaug River and French River valleys north of their confluence. Initial ponding was behind head of Putnam deposits (pt) in Quinebaug valley to the south. Southernmost deltaic surfaces are at 105 m (345 ft) and rise to 117 m (385 ft) in northern part of unit; fluvial feeder deposits are as high as 136 m (445 ft). (Webster, Putnam)

ew **East Woodstock deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits in Little River valley; southern sequences have surface altitudes at 105 to 111 m (345 to 365 ft), initially ponded behind high-level uncorrelated unit (u) in valley to the south before its collapse. Northernmost sequences are fluviodeltaic with fluvial feeders reaching 139 m (455 ft) and deltaic surfaces at 117 m (385 ft). (Putnam, Webster)

pt **Putnam deposits**—Southern half of unit consists chiefly of successive ice-marginal deltas in Quinebaug River valley, dammed initially behind generally coarser Danielson deposits (da); delta surfaces rise from 78 m (255 ft) to 84 m (275 ft). Successive ice-marginal fluviodeltaic deposits in northern part of unit have delta surfaces that rise from 87 m (285 ft) to 107 m (350 ft) and fluvial feeder deposits that reach 123 m (405 ft). (Putnam, Danielson)

da **Danielson deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits. Earlier deposits of this unit have deltaic surfaces at 69 to 75 m (225 to 245 ft); topset beds aggraded 3 to 6 m (10 to 20 ft) higher than those of glacial Lake Quinebaug (lqb) just to the south. These earlier deposits of unit da blocked the Quinebaug River valley, leading to construction of later deltas, mostly along Fivemile River, with surfaces at 81 to 84 m (265 to 275 ft). Fluvial terraces reach 87 m (285 ft) at Danielson and 91 m (300 ft) at northern end of unit. Southwest of Danielson, deposits cross till area and are locally graded to low spillways on till. (Danielson)

qb **Quinebaug River valley deposits**—Ice-marginal deltas and fluviodeltaic deposits successively ponded in the Quinebaug River valley initially behind glacial Lake Quinebaug deposits (lqb) at slightly higher levels than projected glacial Lake Quinebaug water plane. Deltas have surface altitudes at 59 to 69 m (195 to 225 ft); several measured altitudes of delta topset-foreset contacts indicate levels of ponding 2 to 6 m (5 to 20 ft) higher than glacial Lake Quinebaug. (Plainfield, Danielson)

smb **Snake Meadow Brook deposits**—Two ice-marginal fluviodeltaic deposits in Snake Meadow Brook valley, a south-flowing tributary to Moosup River. Initial ponding was behind deposits of unit ip, which blocked valley to as high as 102 m

(335 ft) at Almyville. Delta surfaces at southern end of unit are at 96 m (315 ft); fluvial surfaces rise northward to 123 m (405 ft) at ice-margin position where State Route 52 crosses valley. Northern half of unit is ponded behind head of southern sequence; delta surfaces are at 120 m (395 ft) (topset-foreset contact at about 116 m (380 ft)); fluvial surfaces rise northward to 133 m (435 ft) at ice-margin position. (East Killingly, Oneco)

mm **Mount Misery Brook deposits**—Series of four ice-marginal deltaic deposits along Mount Misery Brook between Pachaug River and Lowden Brook; these are succeeded by near-ice-marginal fluviodeltaic sequence north of Lowden Brook. Delta surfaces rise from 78 m (255 ft) to 93 m (305 ft) between Voluntown and Lowden Brook; fluvial sediments in northern part of unit reach 114 m (375 ft). Initially ponded against deposits of adjacent glacial Lake Voluntown (lvo); subsequent ponding behind successive heads were filled with lake-bottom sediments, most of which are now covered by large swamps in valley. (Oneco, Voluntown)

ib **Indiantown Brook deposits**—Predominantly successive ice-marginal deltaic deposits with minor fluvial feeders. Deposits occur in long divide area north of Indiantown Brook, occupied by large kettle-hole ponds. Deltaic surfaces rise from 38 m (125 ft) northward to 50 m (165 ft). Initial ponding was in basin behind bedrock narrows on Shewville Brook, which may have been filled with till or dead ice to an altitude 3 to 5 m (10 to 15 ft) higher than today. Very coarse grained fluvial deposits at northeasternmost ice-margin position reach 59 m (195 ft). (Jewett City, Old Mystic)

cw **Cedar Swamp–Whitford Brook deposits**—Southern part of unit consists of ice-marginal fluvial sediments in Whitford Brook valley and near-ice-marginal fluvial sediments in Williams Brook valley that grade to extensive deltaic deposits to the south. Deltaic deposits have surface altitudes at 44 to 50 m (145 to 165 ft) and are underlain by as much as 24 m (80 ft) of lake-bottom sediment. Fluvial deposits reach 47 m (155 ft) at ice-margin position. Northern part of unit, between Lantern Hill and Cedar Swamp, consists of three successive ice-marginal deltas, including network of esker feeders; last delta heads at segment of Ledyard moraine (hlm), which forms southern border of Cedar Swamp. Initial ponding in this unit was behind head of Mystic–Stonington deposits (lcms) related to glacial Lake Connecticut and a segment of the Old Saybrook moraine (owm). (Old Mystic)

sn **Shunock River deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits in Shunock River valley. Ponding was initially behind head of lower Pawcatuck River deposits (pl) to the south. The first two sequences are ice-marginal deltas with surface altitudes of 20 m (65 ft); north of these, an ice-marginal fluviodeltaic sequence heads at Old Saybrook moraine (owm) position in valley where coarse gravel fluvial sediments have surface altitudes at 41 m (135 ft) and grade southward to deltaic deposits at 26 m (85 ft). North of this ice position, delta surfaces are higher at 44 m (145 ft) to 47 m (155 ft); northernmost deposit in unit is long, near-ice-marginal fluviodeltaic morphosequence heading just downstream from Ledyard moraine (hlm) position with fluvial surfaces as high as 69 m (225 ft). (Old Mystic, Ashaway)

pu **Upper Pawcatuck River deposits**—Successive ice-marginal deltaic and fluviodeltaic deposits in Pawcatuck River valley and its northeast tributaries. Ponding was initially behind part of Shunock River deposits (sn) at Shunock River junction with Pawcatuck River. Earlier sequences are ice-marginal deltas with surface altitudes of 20 to 26 m (65 to 85 ft). Ice-marginal fluviodeltaic deposit heads at segment of Old Saybrook moraine (owm) at southern end of Bell Cedar Swamp; here, fluvial deposits at 38 m (125 ft) slope steeply southward from moraine to delta at 26 m (85 ft). In Ashaway River valley east of there, ice-marginal delta heads at moraine position. North of moraine position in Spaulding Pond, Green Fall River, and Ashaway River valleys, near-ice-marginal fluviodeltaic deposits occur; fluvial sediments as high as 66 m (215 ft) slope steeply southward to deltaic sediments at 32 m (105 ft) ponded behind moraine positions. (Ashaway)

an **Anguilla Brook deposits**—Inferred to be series of dominantly deltaic deposits, initially blocked at narrow bend in valley. Overall gradient of about 3.9 m/km (20 ft/mi) is too steep for simple ponding of one segment behind another and indicates presence of considerable fluvial material. (Old Mystic, Ashaway, Watch Hill)

pl **Lower Pawcatuck River deposits**—At least four successive ice-marginal deltaic and fluviodeltaic deposits in lower Pawcatuck River valley behind Charlestown (hcm) and smaller Avondale (cm) moraines. Overall surface gradient slopes from 41 m (135 ft) southward to 5 m (15 ft). (Ashaway, Watch Hill)

Deposits of Proximal Meltwater Streams—FP Depositional System

Paleogeographic setting: South-draining valleys that had a relatively steep gradient and were not tributary to any glacial lake. These valleys were steep enough to avoid ponding, but not so steep that the sediment load of the meltwater stream was carried beyond. Outwash plains, broad areas of meltwater-stream deposition in front of moraines, also are included.

Deposits: Fluvial sediments in ice-marginal and near-ice-marginal fluvial morphosequences. In general, these deposits are coarse grained (gravel and sand) and only 3 to 9 m (10 to 30 ft) thick (rarely as much as 15 m (50 ft) thick in moraine-proximal outwash deposits).

Stratigraphic arrangement of deposits: Deposits have relatively steep surface gradients, generally on the order of 2 to 7.5 m/km (10 to 40 ft/mi). In Connecticut, deposits of this depositional system are relatively uncommon; the best-developed ice-marginal fluvial deposits occur as parts of fluviodeltaic morphosequences in valleys tributary to glacial lakes and therefore are included in glaciolacustrine systems. Ice-marginal fluvial deposits on outwash plains in front of moraines also are uncommon; only one small deposit is known to occur in Connecticut (part of unit lmwv in the Broad Brook quadrangle), but extensive deposits in front of the Fishers Island-Charlestown moraine occur in the Rhode Island (part of the Watch Hill quadrangle) and in the New York part of the New London quadrangle and are shown on the map.

Map units: Shown in light-orange colors on the map. Unit descriptions (below) include information on identification of valley and topographic situation in which unit occurs; range in thickness of deposits; altitudes of surface slope (gradient), and base-level controls.

fp **Small, uncorrelated proximal fluvial deposits**

Housatonic–Southwest Coastal Basin

nf **Norwalk River–Fivemile River deposits**—Near-ice-marginal fluvial deposits in steeper sections of upper Fivemile River valley and Norwalk River valley. Surface profiles rise from about 14 m (45 ft) in both valleys just north of ice-marginal head of Stamford–Norwalk–Westport deposits (lcsnw) northward to 61 m (200 ft) in Fivemile River valley and 44 m (145 ft) in Norwalk River valley. Composed chiefly of thin (less than 10 m (30 ft)), gravelly, fluvial sediments. Small pockets of ponded sediments may be present locally at depth. (Norwalk North, Norwalk South)

by **Byram River deposits**—Near-ice-marginal fluvial deposits in narrow and relatively steep Byram River valley. Deposits are near sea level at southern end of valley (in New York) and rise to 55 m (180 ft) to the north. Probably constructed as several morphosequences. Small pockets of ponded sediment may be present locally at depth. (Glenville, Mamaroneck)

Naugatuck Basin

nwb **West Branch Naugatuck River deposits**—Ice-marginal and near-ice-marginal fluvial deposits in steep upper reaches of Naugatuck River and its west branch. Northernmost fluvial sequence reaches 30 m (100 ft) and slopes on a steep gradient; southern sequence in city of Torrington slopes from 198 m (650 ft) at northern end to 165 m (540 ft) at southern end, a gradient of about 9.7 m/km (50 ft/mi). (Norfolk, West Torrington, Torrington)

Farmington–Quinnipiac Basin

me Meriden deposits—Fluvial deposits in the Sodom Brook valley grade from boulder-cobble gravel with surface altitude at 59 m (195 ft) at ice-contact head, southward through finer gravel and sand, to sand and some silt underlying Quinnipiac Valley terrace deposits (qt) at South Meriden. Base level for meltwater streams that built this unit may have been glacial Lake Quinnipiac. Character of deposits in Harbor Brook valley in eastern part of unit is uncertain; several well logs from center of Meriden mention sand and silt in subsurface, indicating possible ponding. (Meriden)

Upper Connecticut Basin

ww **Woods Stream–Wrights Brook deposits**—Ice-marginal fluvial deposits in relatively steep, south-draining tributary valleys north of Scantic River; deposited from both sides of an interlobate angle between Connecticut Valley ice lobe and eastern upland ice margin. Heads of deposits in several places lie on drainage divides; ice margin to the north was requisite to deposition in those places. (Broad Brook, Ellington, Springfield South, Hampden)

Lower Connecticut Basin

bu **Upper Blackledge River fluvial deposits**—Series of three near-ice-marginal fluvial deposits in upper Blackledge River valley; steep surface gradients of more than 5.8 m/km (30 ft/mi). (Marlborough, Rockville)

bf **Branford River fluvial deposits**—Series of short, ice-marginal fluvial deposits in Branford River, Pisgah Brook, and Stony Creek. (Branford)

Thames Basin

sbp **Sherman Brook–Bartlett Brook–Pease Brook deposits**—Ice-marginal and near-ice-marginal fluvial morphosequences in steep tributary valleys to Yantic River; surface gradients are 7.8 to 9.7 m/km (40 to 50 ft/mi) and deposits are coarse gravel facies in most proximal parts. (Colchester, Fitchville, Willimantic, Columbia)

jhg **Jordan Brook–Hunts Brook–Green Marsh Brook deposits**—Ice-marginal fluvial deposits in steep sections of three valleys; shingled against proximal fluvial heads of Jordan Cove–lower Thames River deposits (lcjl) in lower part of each valley. Surface gradients in each sequence are 7.8 to 9.7 m/km (40 to 50 ft/mi); first sequence in Hunts Brook valley heads at Ledyard moraine (hlm); sequence in Jordan Brook valley is near-ice-marginal, built from Ledyard moraine position on upland about 0.4 km (0.25 mi) to the north of unit. (Montville, Uncasville, New London, Niantic)

gr **Groton deposits**—Ice-marginal fluvial deposits in steep upper sections of tributary valleys to Poquonock River; head of outwash in Hempstead Brook valley lies on northern divide at 35 m (115 ft) and slopes southward at about 7.8 m/km (40 ft/mi); shingled against ice-marginal fluvial head of Poquonock River deposits (lcp). Deposits in Rosemond Lake valley head at Ledyard moraine (hlm) and also at northern divide at 59 m (195 ft), and slope southward with a gradient of about 4.8 m/km (25 ft/mi) to 32 m (105 ft), shingled against the 41-m (135-ft) head of unit lcp in that valley. (Uncasville, New London)

Quinebaug and Southeast Coastal Basin

kb **Kitt Brook deposits**—Ice-marginal and near-ice-marginal fluvial deposits in Kitt Brook valley. Probably at least three sequences, subtly marked by breaks in surface gradient which, over whole unit, slopes southward from 96 m (315 ft) to 78 m (255 ft). (Scotland, Plainfield)

ca **Carson Brook deposits**—Ice-marginal fluvial deposit in Carson Brook valley; unit is more extensive in Rhode Island. Probably several successive sequences occur within unit. Overall surface gradient slopes southward from 128 m (420 ft) to 102 m (335 ft). Sediment was deposited around numerous large ice blocks. (Oneco)

ao **Avondale outwash deposits**—Small area of ice-marginal fluvial sediments built from Avondale moraine (cm) ice position; unit is more extensive farther east in Rhode Island. (Watch Hill)

cho **Charlestown outwash deposits**—Very coarse fluvial gravel deposits in outwash fans directly in front of Charlestown moraine (hcm); deposits extend below present sea level. Unit shown on map is westernmost end of much more extensive deposits that occur farther east in Rhode Island. (Watch Hill)

fo **Fishers Island outwash deposits**—Coarse fluvial gravel deposits in outwash fan in front of western end of Fishers Island moraine (hcm). May be more extensive below modern sea level but have not been mapped. (New London)

Deposits of Distal Meltwater Streams—FD Depositional System

Paleogeographic setting: South-draining valleys and basins after ice-marginal lakes had drained, allowing distal meltwater to incise, terrace, and redeposit sediment of slightly older ice-marginal meltwater deposits. In some cases, these distal meltwater streams originated at the glacier margin, which was more than 8 km (5 mi) away; in other cases, a glacial lake separated the glacier margin from the site of meltwater-terrace deposition, and the meltwater stream issued from the spillway of a glacial lake.

Deposits: Distal fluvial sediments not traceable to an ice-marginal head. Deposits commonly consist of sand and fine gravel only 1 to 3 m (3 to 10 ft) thick; as much as 9 m (30 ft) thick in more extensive map units. Sediment is commonly lithologically distinct from underlying deposits.

Stratigraphic arrangement of deposits: Deposits occur on terraces erosionally inset into ice-marginal (generally glaciolacustrine) deposits that are older and higher in altitude; these terraces extend for relatively long distances farther down the valleys, and may be entrenched through several higher and older units. These units are the most extensive glaciofluvial deposits in Connecticut.

Map units: Shown in dark-orange colors on the map. Unit descriptions (below) include information on location of particular valley and identification of older units into which fluvial sediments are inset; range in thickness of terrace sediments; specific lithology, if distinct; and discussion of requisite timing and position of ice margin at time of deposition.

fd **Uncorrelated meltwater terrace deposits**

Housatonic–Southwest Coastal Basin

bb **Blackberry River terrace deposits**—Distal-meltwater fluvial-terrace deposits, generally less than 6 m (20 ft) thick, overlying and inset into lake-bottom deposits of glacial Lake Norfolk. Terraces are only about 3 m (10 ft) above present stream levels. Deposited as meltwater continued to flow down Whiting River valley after draining of glacial Lake Norfolk. Part of unit that appears to come from upper Blackberry River is probably of nonmeltwater origin. (Ashley Falls)

Farmington–Quinnipiac Basin

fvt **Farmington Valley meltwater terrace deposits**—Fluvial deposits at 66 to 56 m (215 to 185 ft) in north-flowing section of Farmington River, and at 62 to 59 m (205 to 195 ft) in lower Salmon Brook valley north of Tariffville. Fluvial sand and gravel generally less than 8 m (25 ft) thick overlies deltaic and lake-bottom sediments of glacial Lakes Farmington (lf) and Tariffville (lt) and Simsbury lacustrine deposits (sl). Terraces began to form after Tariffville Gap was opened and meltwater drainage in the Farmington River valley shifted from the southward flow that constructed Quinnipiac Valley terrace deposits (qt) to northward flow through the gap. As lake levels in the upper Connecticut River basin lowered to higher New Britain spillway levels, distal meltwater initially forming fvt terrace discharged through Tariffville Gap and built southwestern part of the Windsor deltaic deposits (lhhw) in glacial Lake Hitchcock. As lake levels continued to drop, the Bradley International Airport delta (lhsb) was deposited into lake by waters constructing the fvt terrace. Meltwater terrace remnants (fd) farther upstream along Farmington River valley in uplands may be correlative with fvt terrace. (New Britain, Avon, Tariffville)

qt **Quinnipiac River valley terrace deposits**—Fluvial deposits inset into earlier ice-marginal deposits of glacial Lakes Farmington and Southington. Head of deposit is at Farmington where surface is at 64 m (210 ft) and extends southward to North Haven where surface is at 11 m (35 ft). In lower Quinnipiac River valley (south of the Quinnipiac Gorge), these sediments overlie lake-bottom deposits of glacial Lake Quinnipiac. In upper Quinnipiac River valley, terrace sediments are inset against and overlie deltaic and lake-bottom sediments of glacial Lakes Farmington and Southington. Sediments contain high percentage of crystalline clasts because they were deposited predominantly by meltwater coming out of western highlands down Pequabuck and Farmington River drainages; this is in contrast to high percentage of Triassic and Jurassic sedimentary clasts in older ice-marginal deposits in Quinnipiac River valley. (Wallingford, Meriden, Southington, New Britain)

Upper Connecticut Basin

hm **Hampden–Somers meltwater terrace deposits**—Connecticut part of unit consists of distal fluvial terrace deposits deposited by meltwater that eroded and redistributed parts of glacial Lake Somers deposits (lso) and contributed sediment to northern parts of Scantic River delta deposits (lhhs) of glacial Lake Hitchcock. Ice-marginal head is several miles to the north in Massachusetts where it consists of ice-proximal fluvial deposits. (Hampden, Ellington)

ht **Hockanum River valley meltwater terrace deposits**—Distal fluvial terrace deposits at 69 m (225 ft), near Vernon-Ellington town line, grade southwest to 47-m (155-ft) altitude of Hockanum River delta deposit (lmh), which was fed by meltwater that built unit ht. Meltwater spilled from glacial Lake Ellington to the north and eroded and redistributed eastern margin delta deposits (lme) as it built unit ht terrace. (Rockville, Manchester)

ft **Farmington River terrace deposits**—Thin fluvial sediments incised into and overlying Bradley International Airport delta deposits (lhsb); deposited by meteoric water of postglacial Farmington River before drainage of glacial Lake Hitchcock. This fluvial terrace is on grade with post-stable-level delta of glacial Lake Hitchcock (lhf). (Windsor Locks)

Glacial Ice-Laid Deposits (Pleistocene, Late Wisconsinan to Illinoian)

Till Deposits

Silty sand, clayey silt-sand, and sandy or silty diamict sediment as matrix, containing 5 to 40 percent pebbles, cobbles, and boulders; generally nonstratified. Large boulders within and on surface of tills range from sparse to abundant. Gravel clasts, boulders, and sand grains in till matrix are subangular to subrounded, generally unweathered. Surface till deposits include two facies: (1) compact, massive till with subhorizontal fissile structure, subvertical jointing, and strongly preferred orientation of long axes of elongate gravel clasts; inferred to be subglacial till of lodgement or basal meltout origin; overlain locally by (2) noncompact, sandier, commonly layered till with minor lenses of sorted sand and gravel; less common masses of laminated, fine-grained sediments and clusters of cobbles and boulders; inferred to be a supraglacial till of meltout origin. Compact till common on lower slopes and locally in areas of bedrock outcrops; noncompact till either forms discontinuous, thin veneer overlying compact till and bedrock, or forms small moraines.

Color, texture, and composition of tills related closely to underlying and northerly adjacent bedrock units (see figure 3 on sheet 2) from which till was derived. Tills in highlands are light-gray sand to silt-sand, containing clasts of gneiss, schist, granitic rocks, minor quartzite, and local mafic rocks. Tills in Central Lowland are dark-reddish-brown to yellowish-brown, silty to clayey silt-sand containing clasts of sandstone, basalt, diabase, and erratic clasts of gneiss, schist, and quartzite. In valleys underlain by marble in western part of State, tills are silt-sand with calcareous matrix containing clasts of marble, quartzite, schist, and gneiss.

Two tills deposited during two separate glaciations occur in superposition in Connecticut (Pessl and Schafer, 1968). Upper till (late Wisconsinan) is most extensive, commonly exposed in shallow surface excavations, especially in areas where till thickness is less than 4 to 5 m (12 to 15 ft). Lower till was deposited during earlier glaciation (probably Illinoian). Lower till is less areally extensive than upper till, is primarily a subsurface unit, and generally is overlain by upper till. Lower till does, however, constitute bulk of material in areas where till thickness is greater than 4 to 5 m (12 to 15 ft); described in thick till (tt) description below. Where both tills are exposed, base of upper till truncates weathered surface of older till. Lower part of upper till commonly displays zone of shearing and brecciation in which clasts of lower till are mixed and incorporated into upper till.

t **Thin till deposits**—Mapped in areas where till generally is less than 4 to 5 m (12 to 15 ft) thick. Discontinuous on slopes or in areas of moderate local relief where bedrock outcrops are numerous and where bedrock surface topography controls local relief of land surface. Predominantly upper till that is loose to moderately compact, generally sandy, commonly stony. Both lodgement and ablation facies present in places

tt **Thick till deposits**—Mapped in areas where till generally is more than 4 to 5 m (12 to 15 ft) thick, in glacially smoothed landforms that mask bedrock surface topography. In places, particularly in drumlins, till thickness exceeds 30 m (100 ft); maximum reported thickness is about 61 m (200 ft). Lower till constitutes bulk of till deposits in these areas, although upper till generally is present at surface. Lower till is moderately to very compact, and is commonly finer grained and less stony than upper till. Oxidized zone (lower part of soil profile developed during period of interglacial weathering) generally is present in upper part of lower till section; this zone commonly shows closely spaced joints stained with iron and manganese oxides

End Moraine Deposits

Composed variably of sandy ablation till, sediment-flow and colluvial materials, bodies of stratified sand and gravel, and dense concentrations of surface boulders. Deposits occur in narrow zones that trend east-northeast, principally in southeastern part of the State and in Long Island Sound. Thicknesses on land range from 3 to 18 m (10 to 59 ft); submerged moraine in seismic records as much as 40 m (131 ft) thick. Moraine landforms include low, smooth, or undulating ridges with bouldery surfaces; hummocky, irregular ridges of bouldery till; and dense linear concentrations of boulders without interstitial matrix. In upland areas, morainal segments overlie hills and valleys with little inflection at topographic features. In larger valleys and in Long Island Sound, morainal sediments are associated with heads of meltwater deposits, commonly in arcuate arrays related to position of local topographic lowland. Morainal sediments accumulated at margins of glaciers in zones of stagnant ice, mainly by sediment flow and colluvial processes as ice melted.

m **End moraine deposits, uncorrelated**—Morainal deposits in short, discontinuous segments. Includes narrow morainal ridge at Windsorville village and other small patches of morainal deposits

hcm **Harbor Hill–Fishers Island–Charlestown moraine deposits**—Thick and continuous morainal deposits on Long Island, N.Y. (from Fuller, 1914), Fishers Island, N.Y. (from Goldsmith, 1962; and Upson, 1971), and Rhode Island (from Schafer, 1965). Also includes submerged morainal deposits mapped from seismic lines (see description of unit osm below) in lower lying areas east and west of Fishers Island. This massive morainal deposit accumulated along ice margin during major stand of late Wisconsinan ice sheet approximately 20 km (12 mi) north of terminal moraine. Although unit occurs entirely outside of Connecticut, it is shown on map because these deposits provided dam for glacial Lake Connecticut; deep notch across submerged part of moraine at The Race (just west of Fishers Island) was glacial-lake spillway

cm **Clumps–Avondale moraine deposits**—Clumps moraine segment is chain of small bouldery islands and shoals extending about 12 km (7 mi) from island of South Dumpling, N.Y., to Napatree Point, R.I. Submerged deposits exhibit internal reflection characteristics typical of bouldery till. Submerged ice-marginal lacustrine fan deposits (lcf) in places adjacent to morainal position. Avondale moraine segment is in Rhode Island and consists of narrow ridges of ablation till 6 to 15 m (20 to 49 ft) thick at head of Avondale outwash deposits (ao). Clumps–Avondale moraine is parallel to and about 1 km (0.6 mi) north of Fishers Island–Charlestown moraine. (New London, Mystic, Watch Hill)

mm **Mystic moraine deposits**—Double linear belt of discontinuous, aligned ridges, many of which are extremely bouldery. Moraine extends 20 km (12 mi) from Pine Island south of Groton to Pawcatuck; line of submerged ice-marginal lacustrine fans (lcf) offshore of Groton occurs adjacent to and along trend of Mystic moraine. (New London, Mystic, Watch Hill)

owm **Old Saybrook–Wolf Rocks moraine deposits**—Extends from shoreline at Cornfield Point in Old Saybrook along N. 70° E. trend for about 53 km (33 mi) to Rhode Island border in North Stonington; a few moraine segments were mapped in Rhode Island along same trend to Wolf Rocks in North Kingston. Double linear belt of aligned segments, including many long boulder belts. Individual segments in well-developed double line zone between mouth of Connecticut River and New London are up to 3 km (1.8 mi) in length; these segments mostly stand at heads of ice-marginal deltaic deposits built into glacial Lake Connecticut. Offshore between Old Saybrook and Lordship, submerged extensions become co-extensive with lobate ice-marginal lacustrine fan deposits (lcf). (Essex, Old Lyme, Niantic, New London, Uncasville, Old Mystic, Ashaway)

hlm **Hammonasset–Ledyard moraine deposits—**Double linear belt of aligned segments. Hammonasset belt includes paired, parallel segments composed of linear bodies of till up to 2 km (1.2 mi) in length. Extends 10 km (6 mi) from shoreline areas at Hammonasset Point to village of Westbrook; many segments are at heads of ice-marginal deltaic deposits built into glacial Lake Connecticut. Extends from Hammonasset Point westward beneath Long Island Sound as linear belt of small shoals and islands formed by crest of submerged morainal deposits. East of Connecticut River, extends 36 km (22 mi) from east of Rogers Lake in Old Lyme through East Lyme, Waterford, and Ledyard to Ashwillet Brook in North Stonington; through this section, composed of aligned segments individually up to 3 km (1.8 mi) in length. In valleys, segments are bouldery till bodies at heads of deltaic morphosequences; across upland areas, segments are linear concentrations of large boulders. Probably correlates with several discontinuous morainal deposits in western Rhode Island (Queens River moraine). (Clinton, Essex, Old Lyme, Hamburg, Montville, Uncasville, Old Mystic)

mom **Madison–Oxoboxo moraine deposits**—Madison belt contains six segments of ridges and mounds of bouldery till, individually less than 1 km (0.6 mi) in length with paired, parallel segments at east end. Extends 8 km (5 mi) from Hogshead Point in Guilford eastward to Hammonasset River; offshore to the west, includes three shoals and submerged deposits. Oxoboxo belt east of Connecticut River contains eight short and discontinuous linear patches of large boulders and mounds of hummocky till. Extends 10 km (6 mi) from Barnes Reservoir to Stony Brook Reservoir. (Guilford, Clinton, Essex, Montville)

cnm **Captain Islands–Norwalk Islands moraine deposits**—Two linear belts of aligned emergent islands and shoals, and other submerged deposits. Captain Islands belt includes three islands and peninsula at Greenwich Point in exposed segments 0.08 to 1.20 km (0.05 to 0.75 mi) in length. Extends 10 km (6 mi) from Mansuring Island, N.Y., to Greenwich Point. Norwalk Islands belt contains 13 small islands and numerous shoals composed of nonsorted sediments and subordinate stratified materials exposed in island sea cliffs. Submerged morainal ridge is 10 km (6 mi) long, as much as 2.25 km (1.4 mi) wide, and as much as 17 m (56 ft) high above sea floor. Appears to correlate with Lordship proximal fan deposits (lcf) and interlobate ice-marginal delta position at Lordship (lcl) and continues the submerged proximal fan deposits to Old Saybrook–Wolf Rocks moraine deposits (owm) onshore at Cornfield Point. (Mamaroneck, Bayville, Stamford, Norwalk North, Sherwood Point)

osm **Submerged moraine deposits**—Form positive-relief features independent of underlying topography and lie along trend of emergent moraine lines. Internal seismic reflectors are chaotic; where moraine deposits crop out on sea floor, commonly exhibit parabolic reflectors that represent boulders. Although unit was mapped very conservatively and additional deposits may well exist, six moraine positions previously mapped on land have been extended offshore; these include Harbor Hill–Fishers Island–Charlestown moraine (Schafer and Hartshorn, 1965; Sirkin, 1982) and five recessional positions to the north, including Clumps–Avondale, Mystic, Old Saybrook, Hammonasset–Ledyard, and Madison–Oxoboxo moraines

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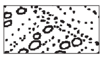
Appendixes 1–3

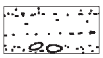
Explanation of Symbols for Appendix 1

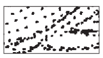
 Boulder gravel, planar bedded, massive, contains imbricate clasts

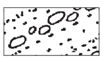
 Pebble-cobble gravel, planar bedded, massive

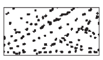
 Pebble-cobble gravel and sand, planar-tabular crossbedded

 Pebble gravel and sand, trough crossbedded

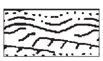
 Coarse sand, planar bedded

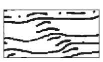
 Coarse sand, trough crossbedded

 Pebble-cobble gravel and sand in planar foreset beds

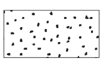
 Coarse sand in trough foreset beds

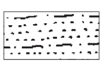
 Coarse sand in trough crossbeds

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

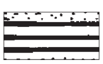
 Fine-medium sand in type B ripple cross laminations, which contain dipping stoss-side and lee-side laminations

 Fine sand and silt in draped laminations

 Fine-medium sand, massive

 Silt and very fine sand in parallel laminations

 Clay laminations

 Fine sand and silt in normally graded lamination

 Fine sand in solitary ripple cross laminations

 Dropstone

 Dropclot of nonsorted material

 Till

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

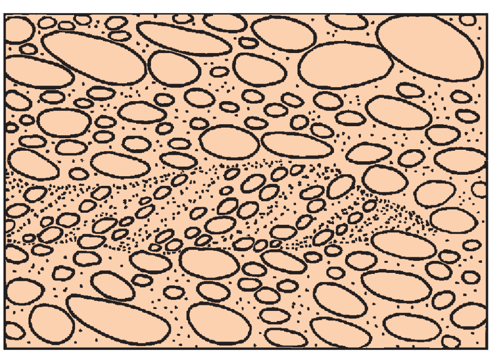
By Byron D. Stone and Janet Radway Stone

Sedimentary Facies

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

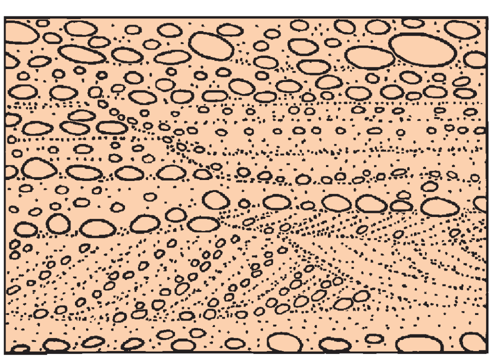
Glaciofluvial Sedimentary Facies

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



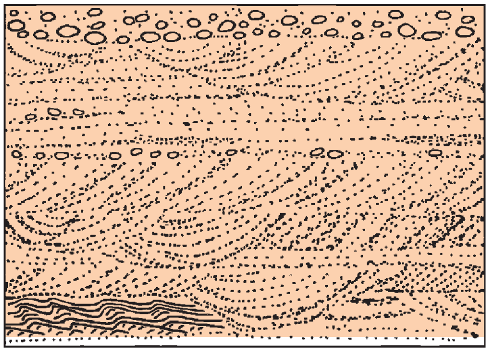
COARSE GRAVEL FLUVIAL FACIES

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



SAND AND GRAVEL FLUVIAL FACIES

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



COARSE PEBBLY SAND FLUVIAL FACIES

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

SAND AND GRAVEL ICE-CHANNEL FLUVIAL FACIES

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

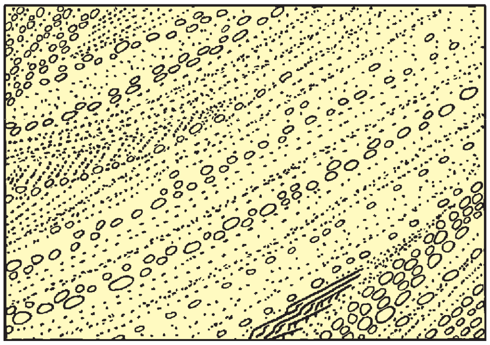
Glaciodeltaic Sedimentary Facies

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

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

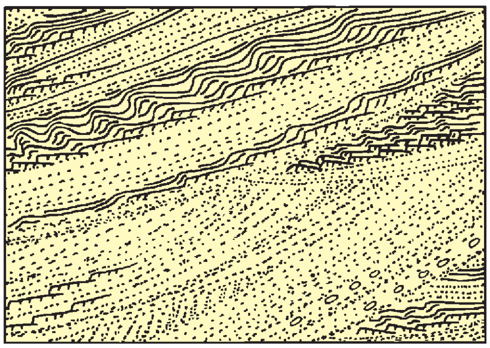
Delta foreset beds are steeply dipping (15°-30°) and are grouped in disconformable sets, commonly 2 to 20 ft thick, which may extend tens of feet down dip. Sets of foreset beds are planar tabular in proximal parts of deltas and trough-shaped in distal parts. Foreset strata contain planar, crossbedded,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

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



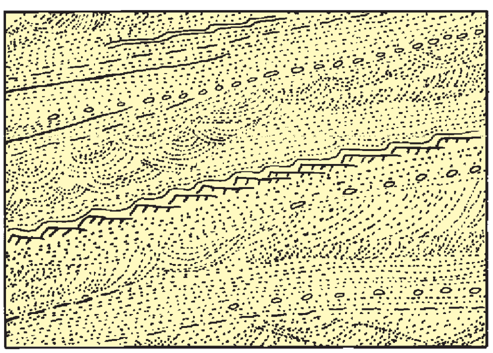
SAND AND GRAVEL DELTAIC FORESET FACIES

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



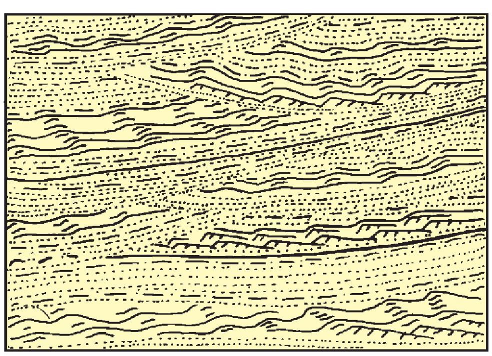
SANDY DELTAIC FORESET FACIES

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



SAND AND GRAVEL DELTAIC BOTTOMSET FACIES

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

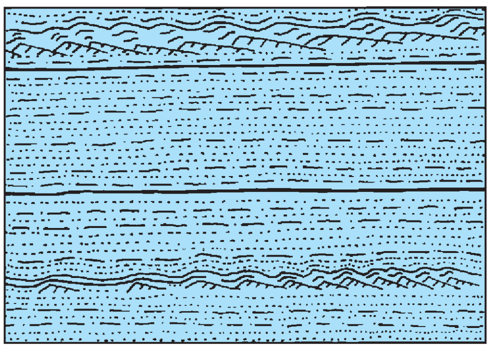


FINE SAND DELTAIC BOTTOMSET FACIES

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

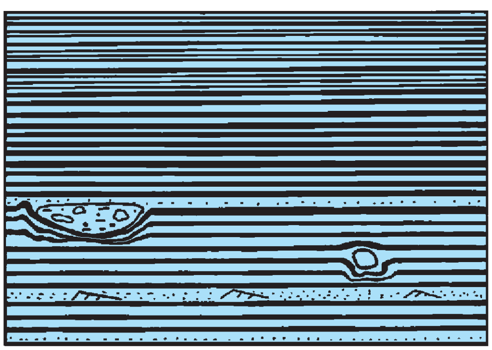
Glacial Lake-Bottom Sedimentary Facies

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



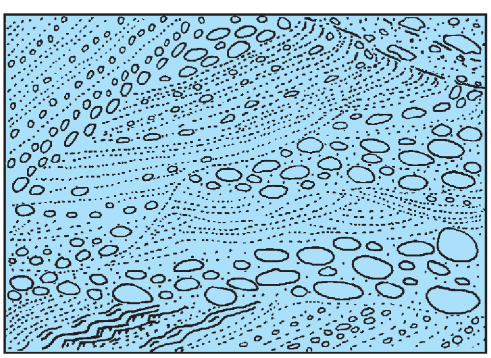
SANDY LAKE-BOTTOM FACIES

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



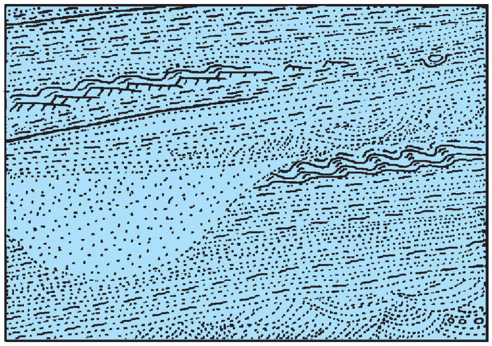
SILT-CLAY LAKE-BOTTOM FACIES

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



SAND AND GRAVEL LACUSTRINE FAN FACIES

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



SAND-SILT LACUSTRINE FAN FACIES

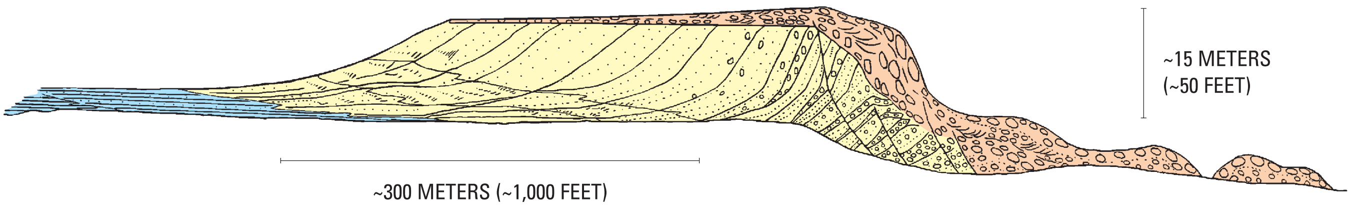
Fine-medium sand in planar and trough bottomset beds, parallel bedded, ripple laminated
Interbedded sets of draped silt and minor clay laminae
6 to 60 ft total thickness
In distal parts of lacustrine fan deposits

Morphosequences

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

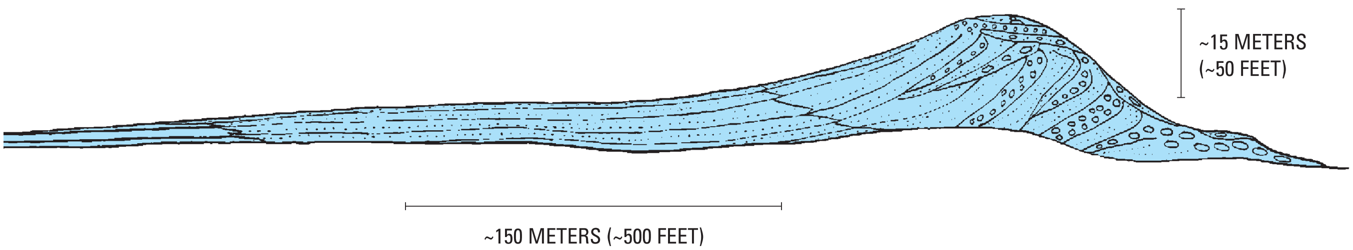
In any one valley, individual morphosequences were deposited sequentially as the ice margin retreated systematically northward. Consequently, in many places the distal, finer grained facies of a younger morphosequence stratigraphically overlies the proximal, coarse-grained facies of a preceding morphosequence.

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



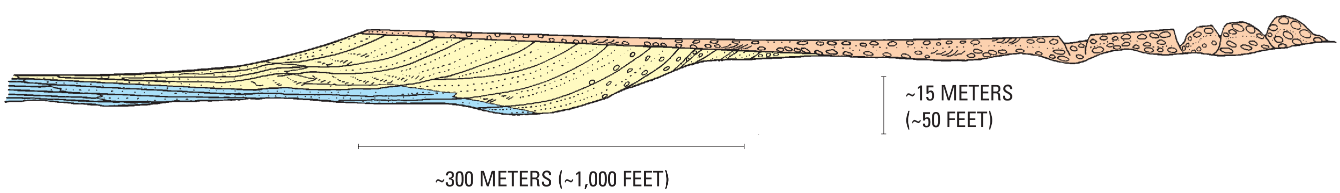
Ice-Marginal Deltaic Morphosequence

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



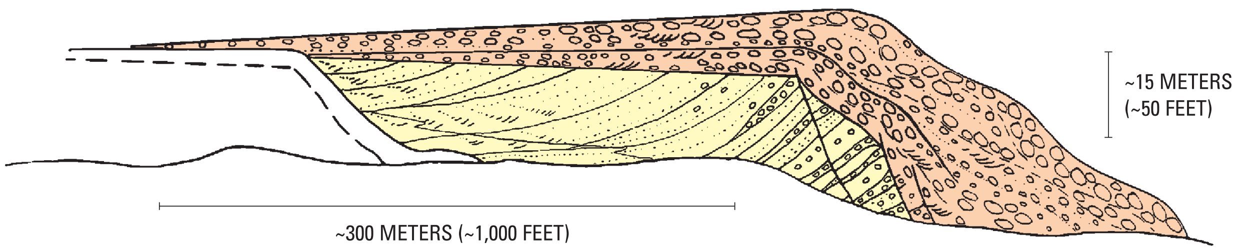
Ice-Marginal Lacustrine Fan Morphosequence

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



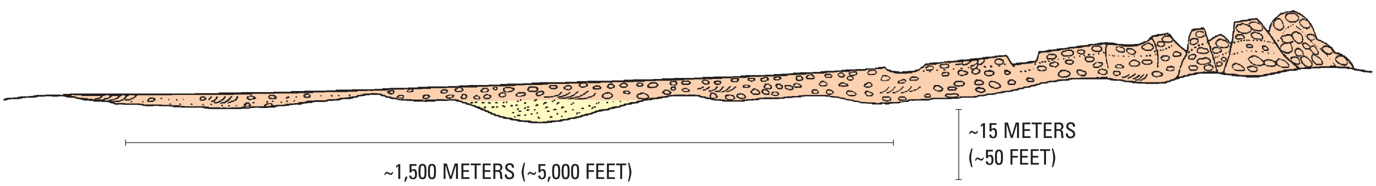
Ice-Marginal Fluviodeltaic Morphosequence

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



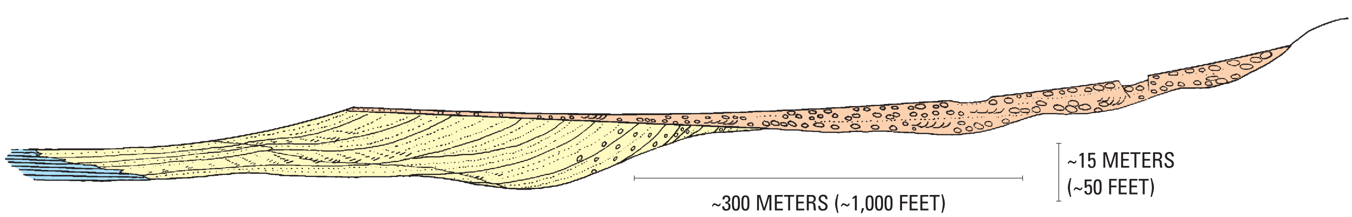
Ice-Marginal Deltaic-Fluvial Morphosequence

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



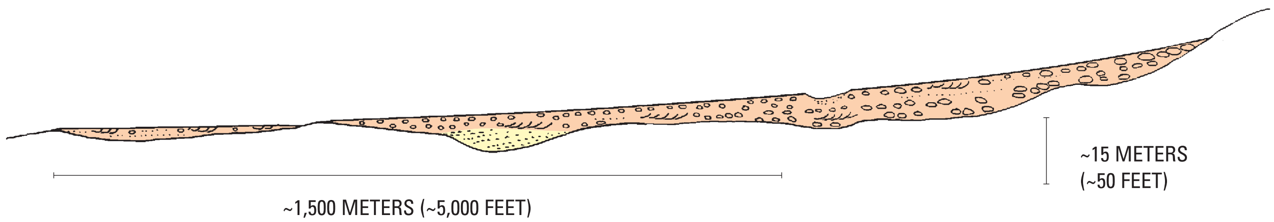
Ice-Marginal Fluvial Morphosequence

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



Near-Ice-Marginal Fluviodeltaic Morphosequence

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



Near-Ice-Marginal Fluvial Morphosequence

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

Other Deposits

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

Lake-Bottom Deposits

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

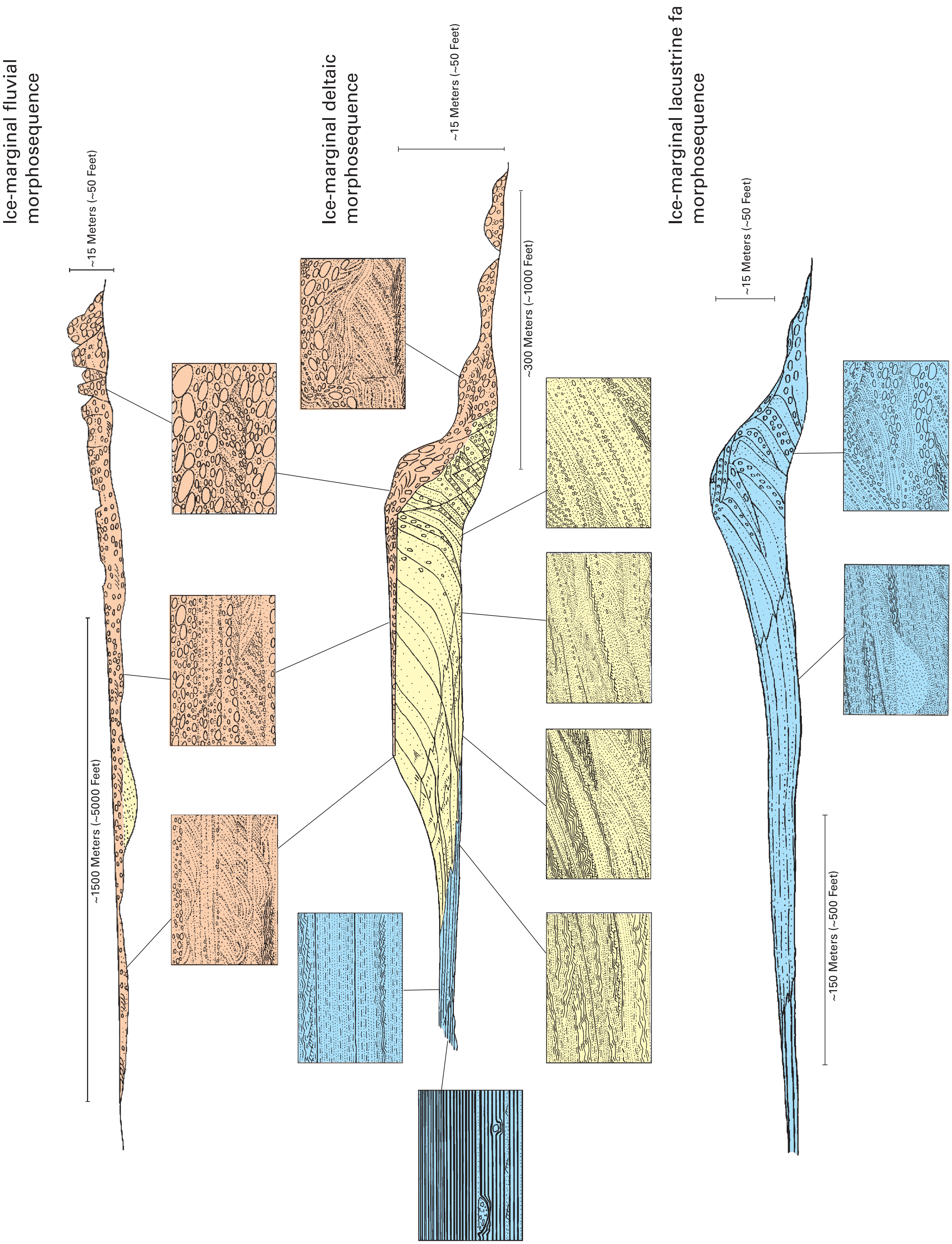
Meltwater Terrace Deposits

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

Meteoric Fluviodeltaic Deposits

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

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



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

Appendix 2. Radiocarbon Dates

Localities 1 to 5 are in western map area, 6 to 21 are in central map area, and 22 to 25 are in eastern map area; localities in each area are numbered from north to south. Quadrangle in which locality occurs is given in parentheses at end of description. References cited are those in which the dates were published and (or) in which significance of the date is discussed in more detail and more exact locations are given.

1 >40,000 B.P. (W-2615) (W.S. Newman, Queens College, written commun., 1981) on peat ball in East Canaan delta of glacial Lake Norfolk; peat contained Cretaceous pollen (N.O. Frederiksen, U.S. Geological Survey, written commun., 1981). Cretaceous organic materials probably derived from deeply buried position in Housatonic valley to the northwest, lowered from originally higher position by solution of marble bedrock (Ashley Falls)

2 12,750±230 B.P. (RL-245) (Kelley, 1975) on bog deposits; provides minimum age for deglaciation of the area (Ellsworth)

3 12,460±110 B.P. (WIS-1405) and 5 younger dates (Gaudreau and Webb, 1985) on lake sediments of Mohawk Pond; provides minimum age for deglaciation of the area (Cornwall)

4 10,190±300 B.P. (W-3931) (Moeller, 1980) and 10,215±90 B.P. (AA-7160) (McWeeney, 1994) on charcoal at “Templeton” paleoindian site at Washington Depot (oldest dated archaeological site in Connecticut) (New Preston)

5 12,880±540 B.P. (GX-9362) (Patton, 1988) on organic material from base of Shepaug River flood-plain alluvium overlying glacial gravels; provides minimum age for deglaciation of the area (Roxbury)

6 12,200±350 B.P. (W-820) (Colton, 1960) on wood from base of organic section of pingo-scar depression in lake-bottom sediments of glacial Lake Hitchcock; provides minimum age for glacial Lake Hitchcock drainage (Stone and Ashley, 1989; Koteff and others, 1988) (Windsor Locks)

7 11,485±115 B.P. (AA-7154) (L.J. McWeeney, Yale University, written commun., 1993) on detrital charcoal (*Pinus* sp.) contained within a dune, indicating that vigorous eolian activity continued until at least this time (Stone and Ashley, 1992) (Springfield South)

8 14,330±430 B.P. (Beta-35211) on detrital wood fragments (probably willow) in eolian sand fill of pingo-scar depression, 12,090±110 B.P. AMS date on single spruce needle in peat, and 11,890±110 B.P. (Beta-34820) on peat at base of organic section fill of pingo-scar depression (Stone and others, 1990). Younger dates support dates from localities 6 and 9 for glacial Lake Hitchcock drainage before the 12.5 to 12.0 ka timespan (Stone and Ashley, 1992) (Hartford North)

9 12,630±240 B.P. (Beta-46514) and 12,320±110 B.P. (Beta-46513) (Robert Gelinas, Ebasco Services, Inc., written commun., 1990) on lower peat section from pingo-scar depression (F7); along with sites 6 and 8, provides minimum age on drainage of southern glacial Lake Hitchcock (Stone and Ashley, 1992) (Hartford North)

10 13,735±180 B.P. (W6397) and 14,120±90 (Beta-52711) (Stone and Ashley, 1992); bulk dates on single sample of plant fragments from upper lake-bottom section of glacial Lake Hitchcock at toe of post-stable stage delta (map unit lhf); 13,540±90 B.P. (Beta-59094, CAMS-4875) AMS date on single fragment of same material as above (Miller, 1995); this date provides timing of early lowering of southern glacial Lake Hitchcock due to breaching of Rocky Hill dam, but before it was completely drained (Hartford North)

11 10,175±75 B.P. (AA-10918) and 9,900±70 B.P. (AA-10919) on seeds from 175-cm and 148-cm depth, respectively, in core; and 9,310±110 B.P. (Beta-51428) and 7,620±80 B.P. (Beta-51427) on bulk peat from 140 to 145 cm and 105 to 110 cm depth, respectively, in core (L.J. McWeeney, Yale University, written commun., 1993). Cores from organic fill in pingo-scar depression, basal organic materials not dated (Stone and Ashley, 1992) (Manchester)

12 10,650±320 B.P. (Y-251) (Flint, 1956) on spruce log from flood-plain alluvium along Park River; locality is on former lake-bottom of glacial Lake Hitchcock, and date must postdate lake drainage. Formerly, this date in combination with locality 13 date was thought to closely postdate lake drainage. More recent dates from localities 6, 8, and 9 provide firm evidence against this interpretation (Stone and Ashley, 1989; Stone and others, 1991) (Hartford North)

13 10,710±330 B.P. (Y-253) (Flint, 1956) on detrital wood in gravel earlier believed to have been deposited by water spilling from the New Britain spillway of glacial Lake Hitchcock. With date from locality 12, was believed to constrain time of lake drainage. New dates from localities 6, 8, and 9 discount this interpretation (New Britain)

14 12,300±350 B.P. (O'Leary, 1975) from plant debris in alluvium of Dickinson Creek; provides minimum postglacial date for the area (Moodus)

15 >35,000 B.P. (W-518, W-519) and 32,350±4,000 B.P. (Y-451) (Hanshaw, 1962) on wood from sediments underlying late Wisconsinan till deposits; assuming dates are finite and that material is in original position, dates represent interglacial time preceding late Wisconsinan glaciation (Meriden)

16 12,700±280 B.P. (W-46) (Leopold, 1955; Deevey, 1958) from sediments recording spruce zone (A) at Durham Meadows. Oscillation in pollen types of this zone were earlier interpreted to represent glacial re-advance (Leopold, 1956); Davis (1965) dismissed oscillation as an artifact of pollen statistics (Durham)

17 6,819±170 B.P. (Y-843) and 3,560±80 B.P. (Y-1077) (Bloom and Stuiver, 1963) on wood from Quinnipiac River alluvial deposits that underlie 5.5 m of tidal-marsh peat; younger date closely predates arrival of tidal-marsh conditions in lower Quinnipiac valley (New Haven)

18 15,090±160 B.P. (Y446-A) (Deevey, 1958; Davis and others, 1980) date on organic material in basal sediment of bog at Totoket village (see description of unit sw). Minimum date for deglaciation of the area (Branford)

19 Series of dates from marsh along Hammock River ranging from 7,060±100 B.P. (Y-1055) to 3,020±90 B.P. (Y-1175) (Bloom and Stuiver, 1963; van de Plassche and others, 1989) that record sea-level rise. Also, a series of dates ranging from 1,710±60 B.P. (GrN-14518) to 85±45 B.P. (GrN-15556) (van de Plassche, 1991) that record latest sea-level-rise history (Clinton)

20 Series of dates from tidal marshes along the lower Connecticut River ranging from 4,475±90 B.P. (GX-15434) to 450±190 B.P. (GX-12079) (Patton and Horne, 1991) that record submergence of former river flood-plain surfaces and the formation of salt and tidal marshes during the last 4,500 years of sea-level rise (Old Lyme)

21 14,240±240 B.P. (Y-950/51) (Davis and Deevey, 1964) corrected to ~15,000 B.P. (Davis and others, 1980) on basal sediments in Rogers Lake. This is the oldest in a long series of dates controlled by careful palynological study. Along with the basal date from the bog at Totoket village (site 18), provides minimum age for deglaciation of southern Connecticut (Old Lyme)

22 13,290±120 B.P. (Y-447 D) and 10,480±140 B.P. (Y-447E) from basal organic sediment in Red Maple Swamp (Beetham and Niering, 1961). Older date from A-1 pollen zone provides a minimum deglaciation date and records a tundra climate in southern Connecticut at that time; younger date is from A-4 pollen zone (Uncasville)

23 Series of dates from multiple cores (23a-23e) in Cedar Swamp, ranging from 15,210±80 B.P. (Beta-66349, CAMS9260) on dwarf willow twigs and Dryas leaves to 5,740±70 B.P. (Beta-61527) on charcoal and peat (McWeeney, 1994); also includes another date of 12,690±120 B.P. (WIS-1875) from spruce pollen zone sediments in Cedar Swamp (Thorson and Webb, 1991); this study also includes a number of younger dates that calibrate the pollen stratigraphy. Sediments containing herb pollen zone beneath spruce zone are biostratigraphically correlated with 15 ka dates from bog at Totoket village (site 18) and Rogers Lake (Old Mystic)

24 11,160±110 B.P. (WIS-1277) from basal organic sediment from Lantern Hill Pond (Bender and others, 1981). Minimum deglaciation date (Old Mystic)

25 12,455±325 B.P. (GX-18094) (Lewis and others, 1994) on bulk organic material including shells in lower part of 5-m core into marine delta sediments of unit md. These sediments overlie the marine unconformity in Long Island Sound and the unit as a whole records a -40-m relative sea level in the basin. This dated section is near the southern edge and in lower-middle stratigraphic position of the marine delta that is believed from other evidence to have been constructed between 13.5 and 10.0 ka (from Long Island Sound)


OXFORD
WEBSTER
SOUTH-BRIDGE
WALES
MONSON
HAMPDEN
SPRING-FIELD SOUTH
WEST SPRING-FIELD
SOUTH-WICK
WEST GRANVILLE
TOLLAND CENTER
SOUTH SANDIS-FIELD
ASHLEY FALLS
BASHBISH FALLS
THOMPSON
PUTNAM
EASTFORD
WESTFORD
STAFFORD SPRINGS
ELLINGTON
BROAD BROOK
WINDSOR LOCKS
TARIFFVILLE
NEW HARTFORD
WINSTED
NORFOLK
SOUTH CANAAN
SHARON
MILLERTON
EAST KILLINGLY
DANIEL-SON
HAMPTON
SPRING HILL
COVENTRY
ROCKVILLE
MAN-CHESTER
HARTFORD NORTH
AVON
COLLINS-VILLE
TORRING-TON
WEST TORRING-TON
CORNWALL
ELLSWORTH
AMENIA
ONECO
PLAINFIELD
SCOTLAND
WILLI-MANTIC
COLUMBIA
MARL-BOROUGH
GLASTON-BURY
HARTFORD SOUTH
NEW BRITAIN
BRISTOL
THOMAS-TON
LITCHFIELD
NEW PRESTON
KENT
DOVER PLAINS
VOLUN-TOWN
JEWETT CITY
NORWICH
FITCH-VILLE
COLCHES-TER
MOODUS
MIDDLE HADDAM
MIDDLE-TOWN
MERIDEN
SOUTHING-TON
WATER-BURY
WOOD-BURY
ROXBURY
NEW MILFORD
PAWLING
ASHAWAY
OLD MYSTIC
UNCAS-VILLE
MONT-VILLE
HAMBURG
DEEP RIVER
HADDAM
DURHAM
WALLING-FORD
MOUNT CARMEL
NAUGA-TUCK
SOUTH-BURY
NEWTOWN
DANBURY
BREWSTER
WATCH HILL
MYSTIC
NEW LONDON
NIANTIC
OLD LYME
ESSEX
CLINTON
GUILFORD
BRANFORD
NEW HAVEN
ANSONIA
LONG HILL
BOTSFORD
BETHEL
PEACH LAKE
WOOD-MONT
MILFORD
BRIDGE-PORT
WESTPORT
NORWALK NORTH
POUND RIDGE
MOUNT KISCO
SHERWOOD POINT
NORWALK SOUTH
STAMFORD
GLENVILLE
BAYVILLE
MAMA-RONECK
10 0 10 Miles

Appendix 3. Sources of Onshore Data by 7.5-Minute Quadrangle

AMENIA
Thompson, W.B., 1980, unpublished data.

ANSONIA
Flint, R.F., 1968, The surficial geology of the Ansonia and Milford quadrangles, with maps: Connecticut Geological and Natural History Survey Quadrangle Report 23, 36 p.

ASHAWAY
Schafer, J.P., 1968, Surficial geologic map of the Ashaway quadrangle, Connecticut-Rhode Island: U.S. Geological Survey Geologic Quadrangle Map GQ–712.

ASHLEY FALLS
Holmes, G.W., and Newman, W.S., 1971, Surficial geologic map of the Ashley Falls quadrangle, Massachusetts-Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–936.
Warren, C.R., and Harwood, D.S., 1978, Deglaciation ice fronts in the South Sandisfield and Ashley Falls quadrangles, Massachusetts and Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–1016.

AVON
Radway, J.A., and Schnabel, R.W., 1976, Map showing unconsolidated materials, Avon quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–514–C.
Schnabel, R.W., 1962, Surficial geology of the Avon quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–147.

BASHBISH FALLS
Zen, E-An, and Hartshorn, J.H., 1966, Geologic map of the Bashbish Falls quadrangle, Massachusetts, Connecticut, and New York: U.S. Geological Survey Geologic Quadrangle Map GQ–507.

BAYVILLE
Schafer, J.P., 1979, unpublished data.

BETHEL
London, E.H., 1984, Surficial geologic map of the Bethel quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–1519.

BOTSFORD
Stone, J.R., and London, E.H., 1985, Surficial geologic map of the Botsford quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–1524.

Figure 1 (facing page). Index map of Connecticut showing 7.5-minute quadrangle names.

BRANFORD
Brown, C.E., 1974, Map showing unconsolidated materials, Branford quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–560–A.
Flint, R.F., 1964, The surficial geology of the Branford quadrangle, with map: Connecticut Geological and Natural History Quadrangle Report 14, 45 p.

BREWSTER
Thompson, W.B., 1980, unpublished data.

BRIDGEPORT
Barker, R.M., 1984, unpublished data.

BRISTOL
Simpson, H.E., 1961, Surficial geology of the Bristol quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–145.

BROAD BROOK
Colton, R.B., 1965, Geologic map of the Broad Brook quadrangle, Hartford and Tolland Counties, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–434.
Langer, W.H., and Colton, R.B., 1973, Map showing unconsolidated materials, Broad Brook quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–451–B.

CLINTON
Flint, R.F., 1971, The surficial geology of the Guilford and Clinton quadrangles, with maps: Connecticut Geological and Natural History Survey Quadrangle Report 28, 33 p.

COLCHESTER
Stone, J.R., 1981, unpublished data.

COLLINSVILLE
Brown, C.E., and Colton, R.B., 1973, unpublished data.
Colton, R.B., 1970, Preliminary surficial geologic map of the Collinsville quadrangle, Litchfield and Hartford Counties, Connecticut: U.S. Geological Survey Open-File Map.

COLUMBIA
Ziska, M.A., 1978, The surficial geology of the Columbia quadrangle: Storrs, Conn., University of Connecticut, unpublished M.S. thesis, 81 p.

CORNWALL
Warren, C.R., and Colton, R.B., 1974, Surficial geologic map of the Cornwall quadrangle, Litchfield County, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–1148.

DANBURY

Thompson, Woodrow, 1975, Surficial geologic map of the Danbury quadrangle, Connecticut: U.S. Geological Survey Open-File Report 75–547.

DANIELSON

Randall, A.D., and Pessl, Fred, Jr., 1968, Surficial geologic map of the Danielson quadrangle, Windham County, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–660.

DEEP RIVER

O'Leary, D.W., 1973, Map showing unconsolidated materials, Deep River quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–524–A.

O'Leary, D.W., 1977, Surficial geologic map of the Deep River quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–1370.

DOVER PLAINS

Thompson, W.B., 1980, unpublished data.

DURHAM

Simpson, H.E., 1968, Surficial geologic map of the Durham quadrangle, Middlesex and New Haven Counties, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–756.

EASTFORD

Pease, M.H., Jr., 1972, Geologic map of the Eastford quadrangle, Windham and Tolland Counties, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–1023.

EAST KILLINGLY

London, E.H., and Melvin, R.L., 1980, unpublished data.

ELLINGTON

Colton, R.B., 1972, Surficial geologic map of the Ellington quadrangle, Hartford and Tolland Counties, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ–965.

Koza, D.M., 1979, Map showing unconsolidated materials, Ellington quadrangle, Connecticut: U.S. Geological Survey Miscellaneous Field Studies Map MF–604–B.

ELLSWORTH

Kelley, G.C., 1975, Surficial geologic maps of the Kent and Ellsworth quadrangles, Litchfield County, Connecticut: U.S. Geological Survey Open-File Report 75–171.

Kelley, G.C., 1977, Late Pleistocene and Recent geology of the Housatonic River region in northwestern Connecticut: U.S. Geological Survey Open-File Report 77–545, 222 p.

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