GLACIAL GEOLOGY AND GEOMORPHOLOGY OF THE PASSAIC, HACKENSACK, AND LOWER HUDSON VALLEYS, NEW JERSEY AND NEW YORK

Scott D. Stanford
New Jersey Geological Survey
P. O. Box 427
Trenton, NJ 08628
Scott.Stanford@dep.state.nj.us

Introduction

The New York City area is a gateway between glaciated terrain to the north and preglacial terrain and the Atlantic Ocean to the south. The Hudson valley is one of the three principal conduits into the Atlantic for meltwater draining from interior North America. Meltwater and sediment discharges from the Hudson have been invoked as the source of glaciogenic deposits and erosional features on the New Jersey continental shelf. Freshwater outflows from glacial lakes in the Ontario basin discharged down the Hudson during the late Wisconsinan deglaciation. Those outflows may have triggered the Younger Dryas and earlier climate coolings by suppressing ocean circulation in the North Atlantic (Donnelly and others, 2005; Rayburn and others, 2007; Obbink and others, 2010). More locally, glacial deposits are productive aquifers in places. Their strength properties and permeability are important features to consider when engineering infrastructure and foundations, modeling groundwater flow, and remediating groundwater and soil contamination.

Deposits of at least three glaciations are present in the New York City area. The earliest glaciation, known as the pre-Illinoian, may have occurred in the late Pliocene, between 2 and 2.5 Ma. A second glaciation is likely of late Illinoian age (~150 ka, oxygen-isotope stage 6), although it may be somewhat older. For ease of reference, it will be referred to as the Illinoian glaciation in this paper, although an Illinoian age is not proven. The most recent glaciation is of late Wisconsinan age and reached its maximum position, as dated by radiocarbon, between 21 and 20 ka (all dates related to the late Wisconsinan glaciation are stated in radiocarbon years). Fluvial deposits south of the glacial limit, and fluvial erosional features preserved within glaciated terrain, provide evidence for preglacial and interglacial routes of the Hudson and its tributaries. This paper will survey the geomorphic and glacial history of the lower Hudson, Hackensack, and Passaic valleys in northeastern New Jersey and adjacent New York, and will discuss lake-drainage events in the upper Hudson valley that affected the lower valley and shelf during the late Wisconsinan deglaciation. A final section will discuss the glacial stratigraphy of the buried valley aquifer system in the upper Passaic basin.

The observations presented here are based on surficial geologic mapping at 1:100,000 scale in New Jersey conducted in the 1980s and early 1990s as part of a cooperative mapping program by the N. J. Geological Survey and the U. S. Geological Survey (Stone and others, 2002), and subsequent 1:24,000 quadrangle mapping by the NJGS in the 1990s and 2000s conducted in part with funding from the Statemap component of the National Cooperative Geologic Mapping Program, administered by the USGS. All of the quadrangles covering the New Jersey part of the study area have been mapped and published (pdfs for all maps can be viewed and downloaded at (http://www.njgeology.org/pricelst/geolmapquad.htm). These mapping efforts built on the excellent early work of R. D. Salisbury, C. E. Peet, and H. B Kummel (in Salisbury, 1895, 1902; Merrill and others, 1902; Bayley and others, 1914). Observations in New York also draw from mapping by Woodworth (1901) and Fuller (1914) on Long Island, followed by subsurface mapping conducted as part of

groundwater-resource investigations in Brooklyn and Queens (Perlmutter and Geraghty, 1963; Soren, 1978; Buxton and Shernoff, 1999), and the regional map compilation of Cadwell (1989). Lacustrine and fluvial features in the upper Hudson valley discussed in the section on Lake Albany were compiled by Stanford (2010) from the works listed in the caption for fig. 6.
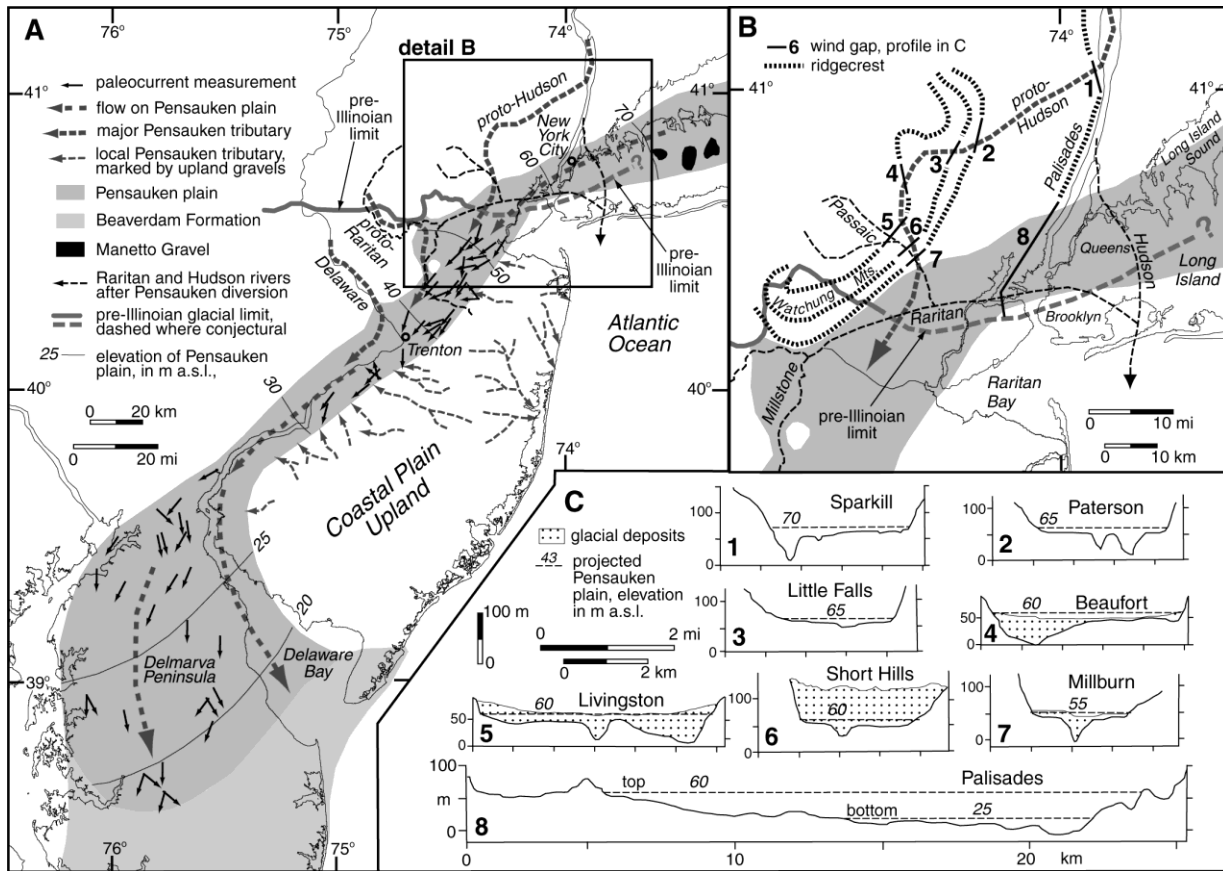
Preglacial Fluvial Drainage

Nonglacial fluvial deposits outside the glacial limit, and wind gaps and buried valleys within the limit, document preglacial and interglacial routes of the Hudson and other rivers in the New York City area. For the purposes of this discussion, drainage before the pre-Illinoian glaciation will be referred to as "pre-early", that before the presumed Illinoian glaciation will be referred to as "pre-intermediate", and that before the late Wisconsinan glaciation will be referred to as "pre-Wisconsinan".

Pre-early drainage is documented by the Pensauken Formation (fig. 1). The Pensauken is a fluvial braidplain deposits of arkosic quartz sand and quartz-quartzite-chert gravel that forms a 10 to 12-mile (15 to 20-km) wide plain along the inner edge of the Coastal Plain from the glacial limit southwestward to the head of the Delmarva Peninsula, where the plain broadens and turns southward (Owens and Minard, 1979). The surface of the plain slopes from an altitude of 165 feet (50 m) near the glacial limit to 65 feet (20 m) in southern Delaware (fig. 1A), where it grades into a marginal-marine sand known as the Beaverdam Formation. Paleoflow direction measured at numerous locations on the tabular, planar cross beds that typify the Pensauken demonstrate southwesterly flow, in agreement with the surface slope. North of the glacial limit the Pensauken is glacially eroded but remnants of it occur beneath till in New Jersey and Staten Island for several miles north of the limit, and distinctive well-rounded white to yellow-stained quartz pebbles from the Pensauken are common in till in the Hackensack lowland and on Long Island. The Manetto Gravel in central Long Island (fig. 1A) is lithically similar to the Pensauken (Suter and others, 1949) and is preserved on uplands outside the glacial limit at elevations similar to those expected if the Pensauken plain were projected northeastward. While the Manetto has been interpreted as a glacial deposit (Fuller, 1914; Sirkin, 1986; Stone and Borns, 1986), its lithology and topographic position suggest that it is a Pensauken equivalent and marks former extension of the plain to the northeast.

A series of wind gaps in the Palisades and Watchung ridges provide further evidence of drainage alignment during Pensauken time. A gap in the Palisades at Sparkill, NY (1 on fig. 1 B and C) and a series of six aligned wind gaps through the three Watchung ridges at, from north to south, Paterson, Little Falls, Beaufort, Livingston, Short Hills, and Millburn (2 through 7 on fig. 1 B and C), all are of similar width and have preglacial rock floors (notched in places by later glacial and local fluvial erosion) that decline from north to south and grade to the Pensauken plain. These gaps mark the route of the Hudson during deposition of the Pensauken (Johnson, 1931; Stanford, 1993). The multiple crossings of resistant basalt and diabase is probably the result of superposition of the Hudson on a covering of Miocene sand (probably inland extensions of the Kirkwood or Cohansey formations of southern New Jersey) in the late Miocene.

A much broader gap is cut into the Palisades Ridge just north of Staten Island (8 on fig. 1 B and C). The 60 and 25 m (200 and 80 ft) elevation lines in this gap bracket the top and bottom of the thalweg of the Pensauken deposit as projected upvalley from the glacial limit. This projection fits the gap and also fits the Manetto remnants to the east. It marks the main trunk of the Pensauken river system, which likely included drainage from southern New England, including precursors to the modern Connecticut and Housatonic rivers. The Hudson was a tributary to this trunk Pensauken river, as was the Delaware, which joined the river to the south at Trenton. An upland in the Coastal Plain to the southeast separated the Pensauken valley from the Atlantic north of Delmarva.

Figure 1. Preglacial and interglacial fluvial drainage, modified from Stanford (2010). A. Pensauken fluvial plain, with paleocurrent measurements (Owens and Minard, 1979; Martino, 1981; Stanford and others, 2002). B. Detail of the New York City area showing location of wind gaps (numbers correspond to profiles on C) and pre-intermediate route of the Hudson, Raritan, and Passaic rivers, marked today by buried valleys, established after the pre-Illinoian glaciation. Location of buried shelf valley from Schwab and others (2003). C. Profiles of wind gaps. Note scale change on profile 8.

The Pensauken in central New Jersey contains pollen indicating a Pliocene age (Stanford and others, 2001), and contains temperate-climate plant fossils (Berry and Hawkins, 1935) that indicate a nonglacial origin. The Beaverdam Formation, with which the Pensauken interfingers downvalley, is also of Pliocene age based on its pollen content (Groot and Jordan, 1999). The Pensauken plain probably aggraded during the mid-Pliocene eustatic highstand centered around 3.5 Ma, although the incised valley in which it aggraded had likely been in place since the latest Miocene, given the depth and extent of erosion into middle and late Miocene upland deposits bordering the Pensauken valley (Stanford and others, 2002).

During or shortly after the pre-Illinoian glaciation in the late Pliocene or early Pleistocene, the Pensauken river was diverted southeasterly in the New York City area, breaching the Coastal Plain upland and initiating direct access of the Hudson to the Atlantic. Details of the diversion are unknown because the depositional record has been removed by subsequent fluvial, marine, and glacial erosion. However, in the Raritan valley just west of the study area, the southernmost deposits of pre-Illinoian till overlie Pensauken Formation remnants and show erosional preservation and weathering intensity similar to the Pensauken, indicating that they are not widely separated in age. It is also clear from the distribution of pre-Illinoian till west of the late Wisconsinan glacial limit that the pre-Illinoian glacier advanced across the Pensauken plain in the New York City area (fig. 1B, see also Stop 3 of field trip).

Following this diversion a new drainage network became established. The Hudson drained southward, probably along the route of what is now the Harlem River and then across Queens by way of a now-buried valley (fig. 1B) (deLaguna, 1948; Soren, 1978; Buxton and Shernoff, 1999). This buried valley extends eastward onto the continental shelf off of Rockaway Beach (Schwab and others, 2003). The lower Raritan, and its tributaries the Passaic and Millstone, established an easterly course on the abandoned Pensauken plain as a tributary to the Hudson, crossing Brooklyn by way of another now-buried valley. Incision of these valleys, to depths of as much as 150 feet below the Pensauken plain, was accomplished in the early and middle Pleistocene, prior to the Illinoian glaciation, because Illinoian glacial deposits, including the Jameco Gravel on Long Island and Illinoian glaciolacustrine deposits and till in the upper Passaic basin, partly fill them. This network constitutes the "pre-intermediate" drainage that was in place before the Illinoian glaciation.

Some of this pre-intermediate drainage was no doubt altered by erosion and deposition during the Illinoian glaciation, although the only record of this alteration is in part of the upper Passaic basin where late Wisconsinan deposits are stacked on top of Illinoian deposits (see discussion of Lake Passaic, below, and fig. 8). Here, the pre-intermediate valleys were overdeepened in places by Illinoian glacial erosion and were then filled, or nearly filled, with Illinoian glacial deposits, indicating that post-Illinoian streams were locally dislocated from their previous routes. The main exit point for the Passaic through Second Watchung Mountain at the Short Hills gap was filled with Illinoian till to an elevation of about 200 feet (60 m), which was sufficient to divert the Passaic northward, in a manner similar to that of today, through the gap at Little Falls, which has a floor elevation of about 180 feet (55 m). Elsewhere, late Wisconsinan glacial erosion has removed all evidence of the post-Illinoian land surface.

Final modifications of the river network were made during the late Wisconsinan glaciation. The present Hudson channel between the Palisades and the Bronx and Manhattan was carved by deep glacial erosion along the outcrop belt of soft Stockton sandstone sandwiched between the resistant Palisades diabase and schist and gneiss of the Manhattan Prong. This erosion created an overdeepened trough more than 350 feet deep. The valley across Queens was filled with till and outwash and blocked by deposition of the terminal moraine. Similarly, the lower Raritan valley in New Jersey and Brooklyn was filled with outwash and blocked by the terminal moraine. The Raritan was diverted southeastward across a low shale divide and cut a narrow, gorge-like valley into the shale between Bound Brook and New Brunswick. East of New Brunswick, the rerouted Raritan entered, and broadened and deepened, a pre-existing valley in Cretaceous deposits that drained eastward to Perth Amboy and then into what is now Raritan Bay (Stanford, 1993).

Pre-Illinoian Glaciation

Pre-Illinoian till, and a few glaciolacustrine deposits, occur west of the late Wisconsinan glacial limit on the Watchung Mountains and intervening valleys (fig. 2). They have been entirely removed by glacial and fluvial erosion to the east, and have not been definitively identified on Long Island, although they surely were present there at one time. In their outcrop area, pre-Illinoian sediments are preserved in erosional remnants on flat hilltops and divides. They are absent from similar flat terrain within valleys below divide levels. This pattern indicates that the valleys were cut into bedrock since the pre-Illinoian glaciation, to depths of between 50 and 150 feet below the pre-Illinoian land surface. Pre-Illinoian deposits are also deeply weathered. Gneiss and arkosic sandstone gravel clasts are fully saprolitized or have thick (>0.5 inch) weathering rinds. Quartzite, quartz, and chert clasts are typically intact but stained; some have thin weathering rinds and are easily fractured with a hammer. Siltstone, quartz sandstone, and shale clasts are intermediate in weathering intensity between the gneisses and quartzites. The till matrix contains much secondary clay from alteration of feldspar in the sand fraction to clay minerals, and is reddish-yellow in color due to accumulation of iron oxides and hydroxides from weathering of mafic minerals.
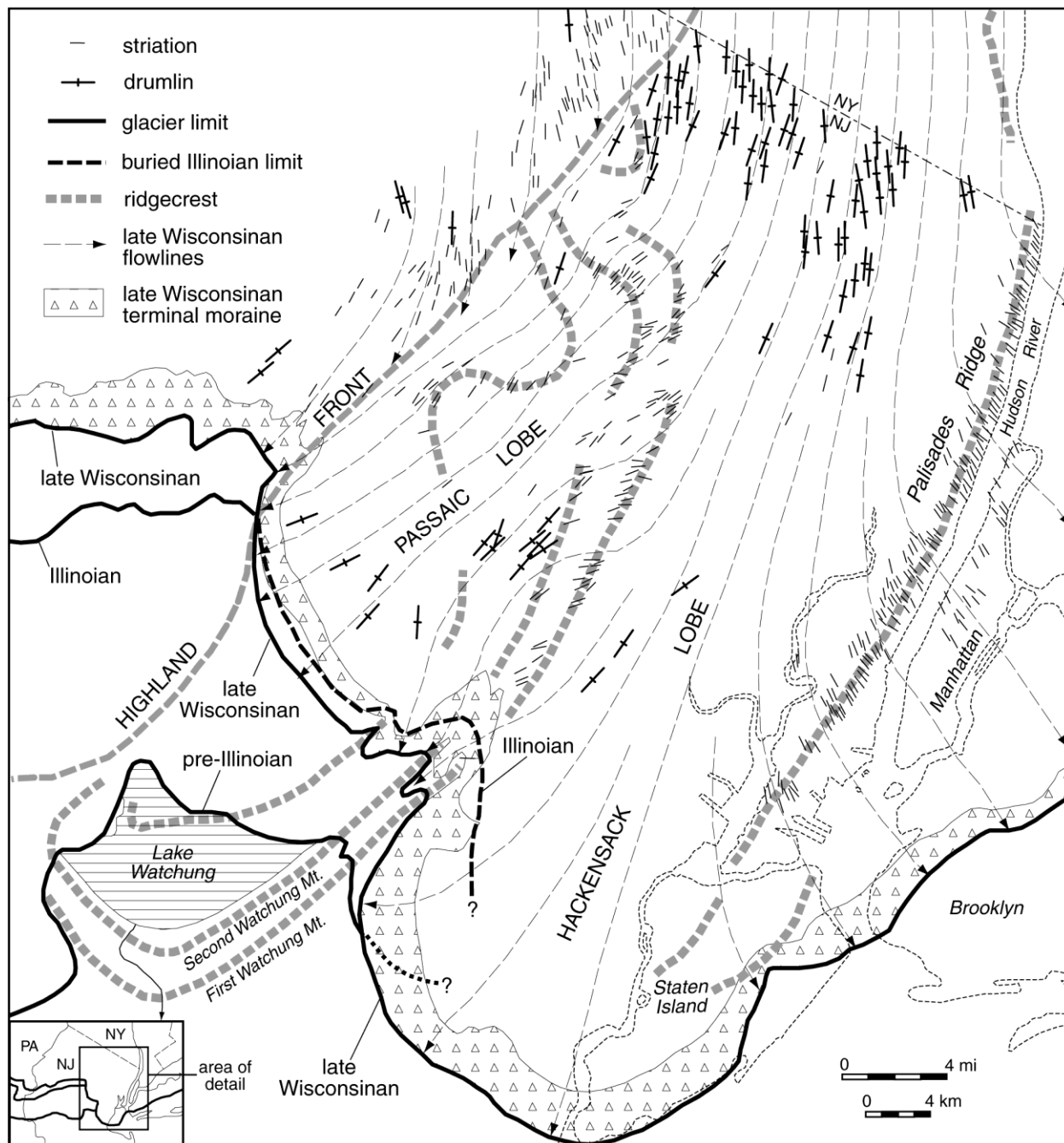
Figure 2. Glacial limits, late Wisconsinan striations and drumlin axes, and late Wisconsinan glacial flowlines. Striations and drumlins are from 1:24,000 surficial geologic maps available at http://www.njgeology.org/pricelst/geolmapquad.htm. Some striations on the Palisades are from Salisbury and Peet (1895). Striations on Manhattan are from Baskerville (1994).

Paleomagnetic measurements on pre-Illinoian lacustrine and fluvial sediments in New Jersey and eastern Pennsylvania yield both normal and reversed signatures, and some reversals are on weathering products (Sasowsky, 1994; Ridge, 2004). While there are many ways to interpret the paleomagnetic data, they at the very least prove a glaciation before 800 ka, the age of the last magnetic reversal. Pollen recovered from basal lake sediments in Budd Lake, a glacially dammed upland basin in the Highlands in western New Jersey, included pre-Pleistocene taxa (Harmon, 1968). Budd Lake lies just outside the

Illinoian and late Wisconsinan glacial limits, and so deposits in the lake basin were not eroded during these glaciations, and long accumulation is possible there. The basin was likely first dammed by deposition of glacial sediments during pre-Illinoian retreat. The Budd Lake pollen, and the similar erosional preservation and weathering properties of the pre-Illinoian till and Pensauken Formation, suggest a Pliocene age for the pre-Illinoian glaciation. The earliest Laurentide glacial deposits are dated by magnetic polarity and volcanic ash in the Missouri River valley to between 2 and 2.5 Ma (Boellstorff, 1978; Roy and others, 2004). The pre-Illinoian deposits here may be an eastern correlate of these earliest mid-continent tills.

The pre-Illinoian limit is mapped based on the well-defined extent of pre-Illinoian till (fig. 2). In addition to the preserved till patches, pre-Illinoian erratics occur within basal parts of the basalt colluvium at the foot of First Watchung Mountain within, but not outside, the limits shown on fig. 2. The limit defines a lobe in the broad Raritan lowland west of the Watchungs and a somewhat less extensive lobe in the Passaic and Hackensack lowlands. The Passaic and Hackensack lowlands, judging from the elevation of the wind gaps discussed in the previous section, were not as deeply eroded in pre-Illinoian time as they are now, or as they were in the middle and late Pleistocene, and so did not serve as effectively as channels for ice as they did in later glaciations. The upper Raritan lowland, in contrast, was largely as it is today and so provided little impediment to pre-Illinoian ice flow.

At its maximum extent the pre-Illinoian glacier enclosed a basin behind Second Watchung Mountain to create glacial Lake Watchung (fig. 2). With the Moggy Hollow gap (which served as the late Wisconsinan lake outlet, see section below on Lake Passaic) closed off by ice, Lake Watchung was controlled by a higher gap to the south that directed lake outflows southward into the Raritan River (Stanford, 2008). A sizable deposit of sand and gravel on the Passaic-Raritan divide at Bernardsville, NJ, is an erosional remnant of a delta deposited in this lake, and smaller deposits of lake clay and lacustrine-fan gravel in the Bernardsville-Basking Ridge area were also laid down in Lake Watchung.

Illinoian Glaciation

Illinoian deposits crop out in a belt several miles wide south of the late Wisconsinan limit west of the Highland Front, and occur beneath late Wisconsinan deposits in the valley fill in the upper Passaic basin between the Highland Front and Second Watchung Mountain (fig. 2, see also section below on Lake Passaic). They have not been observed to the east in the Hackensack lowland but are present in the cores of drumlins to the north near the state line. On Long Island, the Jameco Gravel and Montauk Till, both of which occur beneath Wisconsinan deposits, are likely of Illinoian age.

In contrast to the pre-Illinoian, Illinoian till in the outcrop belt is preserved on gentle and moderate slopes and is eroded only from steep slopes. Illinoian deposits fill modern valley bottoms, and form subdued but recognizable moraines, deltas, and fluvial plains. There has been no incision into bedrock beneath the Illinoian deposits. Gneiss and arkosic sandstone clasts have weathering rinds that are generally <0.5 inch thick and may be partially weathered but are not saprolitized. Matrix color is brown to yellowish-brown rather than reddish-yellow, and the matrix does not contain significant amounts of pedogenic clay. However, in a few places where soil B horizons in Illinoian deposits have been preserved, they are sufficiently well developed to indicate exposure during an interglacial rather than an interstadial (Ridge and others, 1990).

The only radiometric age control for the Illinoian advance is a coral fragment dated by U-Th to 130 ka from a marine sand overlying till on Nantucket (Oldale and others, 1982). This till may be equivalent to the Montauk Till on Long Island, and if so would indicate a pre-Sangamon age for that till, that is, Illinoian (oxygen-isotope stage 6) or older. The degree of weathering also indicates a pre-Sangamon age. On the other hand, the much greater degree of erosion and weathering of the pre-Illinoian

52

deposits indicates a much greater period of time between the pre-Illinoian and Illinoian than between the Illinoian and late Wisconsinan, perhaps an order of magnitude difference.

Late Wisconsinan Glaciation

Advance of Ice

Striation and drumlin orientations show that late Wisconsinan ice advanced across the study area in two lobes, both part of the much larger Hudson-Champlain lobe (fig. 2). The largest mass, the Hackensack lobe, flowed south-southwesterly down the broad Hackensack lowland underlain by shale and sandstone between the Palisades Ridge to the east and First Watchung Mountain to the west. In the axis of the lowland, unimpeded by topography, it extended southward to a terminus at Perth Amboy, NJ, and the southern tip of Staten Island. At 40°30' this is the southernmost glaciated point in North America east of the Ohio River valley. The east sector of the lobe moved southeasterly over Manhattan, Brooklyn, and Queens to the terminal position. The western sector of the lobe flowed southwesterly over First and Second Watchung mountains and into the upper Passaic lowland, underlain by shale, to form the Passaic lobe. Impeded by the Watchungs, the Passaic lobe did not reach as far south as the Hackensack lobe. A third lobe, flowing out of the Wallkill valley, advanced southward across the Highlands to the west of the Highland Front. Impeded by the topographic barrier of the Highlands, it did not advance as far as the other two and did not enter the Passaic lowland (except at the mouth of the Wanaque valley, see Stop 4 of field trip).

During advance the glacier eroded elongate overdeepened troughs into shale and sandstone bedrock in the Passaic and Hackensack lowlands and Hudson valley. These troughs show as much as 300 feet of overdeepening and extend to more than 350 feet below sea level in the New York City area and, in the Hudson valley, as much as 700 feet below sea level at and north of the Tappan Zee (Newman and others, 1969). To the south they shallow and transition into buried pre-Wisconsinan fluvial topography within about 10 miles of the glacial limit. The troughs in the Passaic and Hackensack are parallel to ice flow, suggesting they were produced by scour. The trough beneath the Hudson, which is excavated chiefly in sandstone, is nearly perpendicular to ice flow over the Palisades, suggesting that it may have been produced by plucking like the Palisades cliffs themselves, or perhaps by scour from local basal ice channeled southwesterly along strike of the sandstone belt.

Terminal Moraine

At its maximum extent, the ice front stabilized and built a prominent terminal moraine in a belt averaging about 2 miles (3 km) wide and consisting of till between 40 and 200 feet (12 and 60 m) thick. The moraine belt consists of ridges, knolls, and basins with as much as 100, but generally less than 50, feet (30 and 15 m) of relief. By volume of till, the terminal moraine is at least an order of magnitude larger than the few recessional moraines mapped in the northeastern United States.

Varves deposited in the southern basin of Lake Passaic (fig. 3, portion of Lake Passaic south of margin TM) provide an estimate of residence time at the moraine. This lake basin formed when ice arrived at the terminus and was isolated from glacial sediment input when the ice front retreated from the moraine, because the moraine itself and its fronting delta formed a continuous barrier above lake level along the north edge of the basin. Reimer (1984) counted 750 glacial varves in the basin, overlain by at least 450 much thinner "microvarves" that likely represent postglacial sedimentation. A 750-year duration for residence at the terminal moraine is consistent with regional bracketing radiocarbon dates (fig. 4).

Looping of the moraine at the margin of the Hackensack lobe around the Todt Hill upland underlain by serpentinite bedrock on Staten Island indicates that the glacier surface rose at a rate of about 300 feet per mile, which is the same gradient observed along terminal moraine loops in the New Jersey

Highlands. This rate of rise indicates that, at maximum advance, the late Wisconsinan glacier was between 1500 and 2000 feet thick over Manhattan, somewhat less than claimed in disaster movies but enough to bury the Empire State Building.
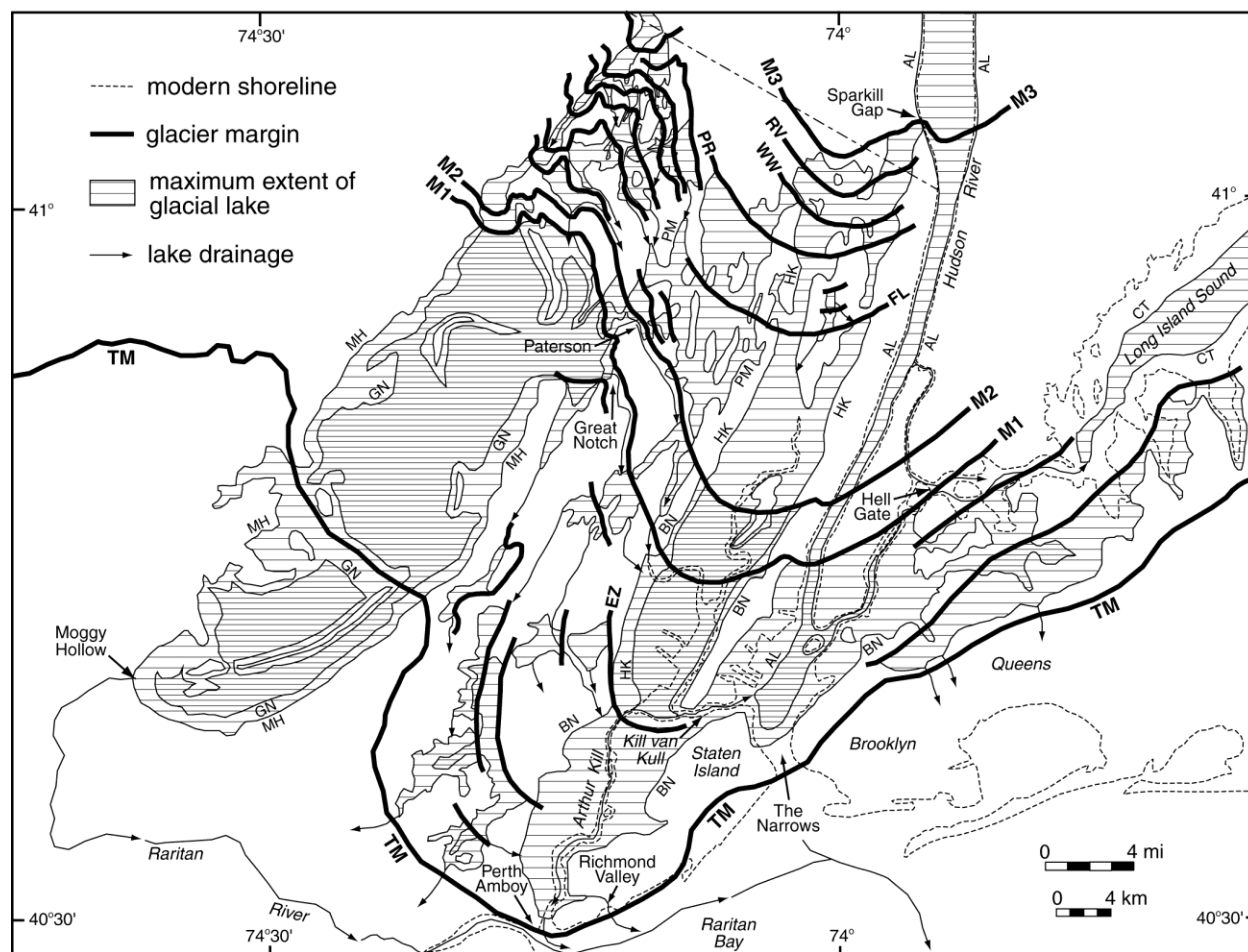
Figure 3. Glacial lakes and ice margins during late Wisconsinan retreat. Lakes are identified by the following abbreviations on their shorelines: AL=Albany, BN=Bayonne, CT=Connecticut, HK=Hackensack, PM=Paramus, MH=Passaic, Moggy Hollow stage, GN=Passaic, Great Notch stage. Ice margins are: TM=terminal moraine, M1=last ice margin before Lake Bayonne lowers to form Lake Albany, Hell Gate stage, in the Hudson valley, and before Lake Passaic lowers from the Moggy Hollow stage to the Great Notch stage, M2=last ice margin before Lake Passaic, Great Notch stage drains, and before spillway erosion establishes stable Lake Hackensack, M3=last ice margin before Lake Hackenssack lowers through Sparkill Gap into Lake Albany. Recessional ice margins marked by large deltas or glaciofluvial plains are: EZ=Elizabeth, FL=Fair Lawn, PR=Paramus, WW=Westwood, RV=Rivervale.

Retreat

Retreat from the moraine is documented by lacustrine stratified deposits, and a few glaciofluvial and ice-contact deposits, that mark ice-margin positions or that record lake stages that in turn require certain ice-margin geometries (fig. 3). Lake-stage elevations are marked primarily by delta-plain elevations or, ideally, by the elevation of the contact of fluvial topset beds and lacustrine foreset beds within deltas, although these contacts are rarely exposed. Discontinuous beach features have been observed in Lake Passaic (Salisbury and Kummel, 1895), and were likely present in other broad lake

basins with sufficient fetch like lakes Bayonne and Hackensack, but today all such fine topographic details have been erased by urbanization.

Lakes formed in valleys that drained toward, and so were dammed by, the glacier margin, or in valleys and lowlands dammed by earlier glacial deposits. In the latter class are Lake Bayonne (BN on Fig. 3) and the Hell Gate stage of Lake Albany (AL on fig. 3), which were dammed by the terminal moraine, and Lake Paramus (PM on fig. 3) which was dammed by an earlier delta that clogged a narrow reach of the lower Passaic valley. In the former class are the many small lakes along the west edge of the Hackensack lobe, which occupied east- or northeast-draining tributary valleys in the Rahway, lower Passaic, and Saddle river valleys that were dammed by the retreating ice front. Lake Passaic is a combination of both types, with an early stage of the lake forming when ice blocked the northeast-draining upper Passaic valley at Paterson, and a maximum (MH on fig. 3) and recessional (GN on fig. 3) stage held in by the terminal moraine dam in the Short Hills gap.

During retreat the terminal moraine dammed the Arthur Kill lowland between Perth Amboy and Staten Island and also the Narrows between Staten Island and Brooklyn, forming the basin occupied by Lake Bayonne. The earliest level of Lake Bayonne was controlled by a spillway across the moraine at Richmond Valley on Staten Island but this was soon succeeded by an eroding spillway across the moraine at Perth Amboy. Lake Bayonne expanded northward to the Newark area and eastward into lower Manhattan as the ice front retreated. When ice uncovered a till upland in northwestern Queens (M1 on fig. 3), a lower spillway draining eastward into the Long Island Sound lowland opened and Lake Bayonne drained. This lower spillway soon stabilized on gneiss bedrock at what is now Hell Gate in the East River at an elevation of -30 feet (-9 m). At this time the Long Island Sound lowland was occupied by glacial Lake Connecticut, which was controlled by a moraine dam at The Race off the east end of Long Island (fig. 5) (Lewis and Stone, 1991). At its west end, the level of Lake Connecticut at this time was -70 feet (-21 m) or lower, providing at least 40 feet (12 m) of drop for the Hell Gate spillway to function.

As Lake Bayonne lowered and drained it was replaced in the Hudson valley by the Hell Gate stage of Lake Albany, controlled by the stable spillway on gneiss in Hell Gate. In the Hackensack valley to the west of the Palisades, outflows from the remnant of Lake Bayonne eroded the Arthur Kill and Kill van Kull channels to uncover Palisades diabase bedrock at Tremley Point in the Arthur Kill at -30 feet (-9 m) and at the west end of the Kill van Kull at -20 feet (-6 m). These became the stable spillways for Lake Hackensack, which was 40 feet (12 m) higher than the Hell Gate stage of Lake Albany.

The uncovering of Hell Gate also coincides with the uncovering of Great Notch in First Watchung Mountain, which caused Lake Passaic to lower 80 feet (24 m) from its highest (Moggy Hollow, MH on Fig. 3) stage to the Great Notch (GN on fig. 3) stage (M1 on fig. 3). This lowering released about 2.5 mi³ (10 km³) of water into Lake Bayonne via the Third River sluice, which provided an erosional pulse that deepened the Arthur Kill and Hell Gate channels. With continued retreat, the Paterson gap in First Watchung Mountain (M2 on fig. 3, Stop 5 on field trip) was uncovered, and the Great Notch stage itself drained, releasing about 1.2 mi³ (5 km³) of water down the Weasel Brook sluice into the lower Passaic valley. This outflow provided a final erosional pulse in the Arthur Kill and Kill van Kull channels, uncovering the spillways for Lake Hackensack and, possibly, completing erosion down to gneiss bedrock at Hell Gate.

Continued retreat uncovered Sparkill gap, a broad wind gap containing a narrow inner notch with a floor at an altitude of 30 feet (9 m) in the Palisades ridge just north of the state line (M3 on fig. 3). Lake Hackensack lowered 40 feet (12 m) to the level of Lake Albany, Hell Gate stage, through the notch. After the lowering, a shallow postglacial lake remained in the northern half of the Hackensack valley. The north end of this postglacial lake was filled with fluvial terrace sand fed by the Passaic and Saddle rivers, which flowed northeasterly across the Lake Paramus basin to Sparkill gap for several thousand years after

deglaciation, until isostatic rebound shifted them to their present southerly flow at about 13 ka (Stanford and Harper, 1991).

During this early postglacial period, before isostatic rebound, the floor of lakes Hackensack and Bayonne in the southern part of the Hackensack lowland was exposed and desiccated. Desiccation created an overconsolidated horizon in the uppermost lake clays, beneath post-rebound alluvial sand and Holocene salt-marsh peat, which is well known from engineering studies (Lovegreen, 1974). Depth of the desiccated zone is as much as 60 feet (18 m) below sea level at Newark and 30 feet (9 m) below sea level at Secaucus which, when adjusted for rebound at the rate of 3.5 feet per mile (0.7 m/km) recorded by delta elevations in stable Lake Hackensack and Lake Albany, Hell Gate stage, approximates the pre-rebound Sparkill gap baselevel at 30 feet (9 m)(Stanford and Harper, 1991).

Chronology

Chronologic control for the late Wisconsinan advance and retreat in the New Jersey-Long Island area is provided by radiocarbon dates and varve counts (fig. 4). All the dates younger than 13 ka are on freshwater peat or terrestrial plant material, except the bone dates for the Sparta mastodon. The older dates are on concretions in varves (the Great Swamp date), and disseminated organic matter in sediment, since no peat or plant macrofossils have been recovered from these deposits, except for the older Long Island date, which is on peat. The two Long Island dates shown on fig. 4 are the youngest of 29 total dates on shells, peat, wood, and organic silt from deformed beds under late Wisconsinan till at Port Washington, Long Island. The other 27 older dates range from 25 to >43.8 ka (Sirkin and Stuckenrath, 1980). Bulk organic sediment and concretions may incorporate old geologic carbon that produces older-than-actual dates for the enclosing deposit. Consistency of the dates, and intervening varve counts, suggests that contamination with old carbon is not a significant problem here. In addition, the dates are consistent with the New England varve chronology. The oldest varves in this chronology date the recessional ice margin at Newburgh, NY to 15 ka (Ridge, 2004, 2008). A minimum of 3200 additional years are recorded by varves in Lake Hackensack and the lower Hudson valley that are not matched to the chronology but precede it (Antevs, 1928). These additional varves indicate that Little Ferry, NJ, the site of the earliest varves, was deglaciated at or before 18.6 ka. This date is consistent with the 19-19.2 ka deglaciation for Little Ferry inferred from the radiocarbon chronology (fig. 4).

Ice arrived at its maximum position at about 21 ka, remained at the terminal moraine for 750-1000 years, based on the glacial varve count in the southern basin of Lake Passaic, and then retreated rather rapidly (an average of 50 m per year, using the smooth recessional curve on fig. 4) up the Hackensack lowland, if the Tappan Zee date of Weiss (1971) is accurate. This date is in organic sediment underlain by 108 feet of varves resting on till. The varves represent a minimum of 200-300 years of elapsed time from deglaciation to deposition of the organic sediment, assuming they are thick (3 to 6 inches or 7 to 14 cm) proximal varves. This elapsed interval is within the one sigma error on the date.

Lake Hackensack varves counted in clay pits at Little Ferry, NJ (Reeds, 1926) show an abrupt thickening at varve year 1097 (Reeds' varve 0) above the basal till. This thickening is likely the result of the lowering of Lake Hackensack when Sparkill gap was deglaciated. At this time, pre-rebound routing of the Passaic and Saddle rivers into the shallow postglacial lake increased sediment inflow, accounting for the thicker varves. The 1097 glacial varves thus measure the duration of retreat from Little Ferry to Sparkill gap, indicating a slower overall retreat rate in the northern Hackensack lowland compared to that further south. This slowed retreat is consistent with the greater volume of ice-contact deltaic sediment laid down in lakes Paramus and Hackensack (from ice margins FL, WW, RV, and M3 on fig. 3) during this interval.

Residence time at these and earlier recessional positions can be estimated from the deglaciation chronology tied to the Tappan Zee date. Large deltas at Newark (in Lake Bayonne), Fair Lawn (in Lake

Paramus), Westwood, Rivervale, and Tappan (all in Lake Hackensack), a glaciofluvial plain and underlying lacustrine valley fill at Elizabeth, and a small recessional till moraine (the Bloomfield moraine) associated with the Newark delta and with other deltaic deposits northwest of Newark, all appear to represent 100- to 300-year intervals of ice-margin stability. Outcrops and test borings showing till over stratified deposits record short (<1 mile) readvances at ice margins EZ (Stanford, 2002) and M1, especially along M1 in the Lake Passaic basin (Salisbury, 1902; Stanford, 2003), where deep water buoyed the ice and increased margin mobility. There is no evidence of readvance elsewhere.
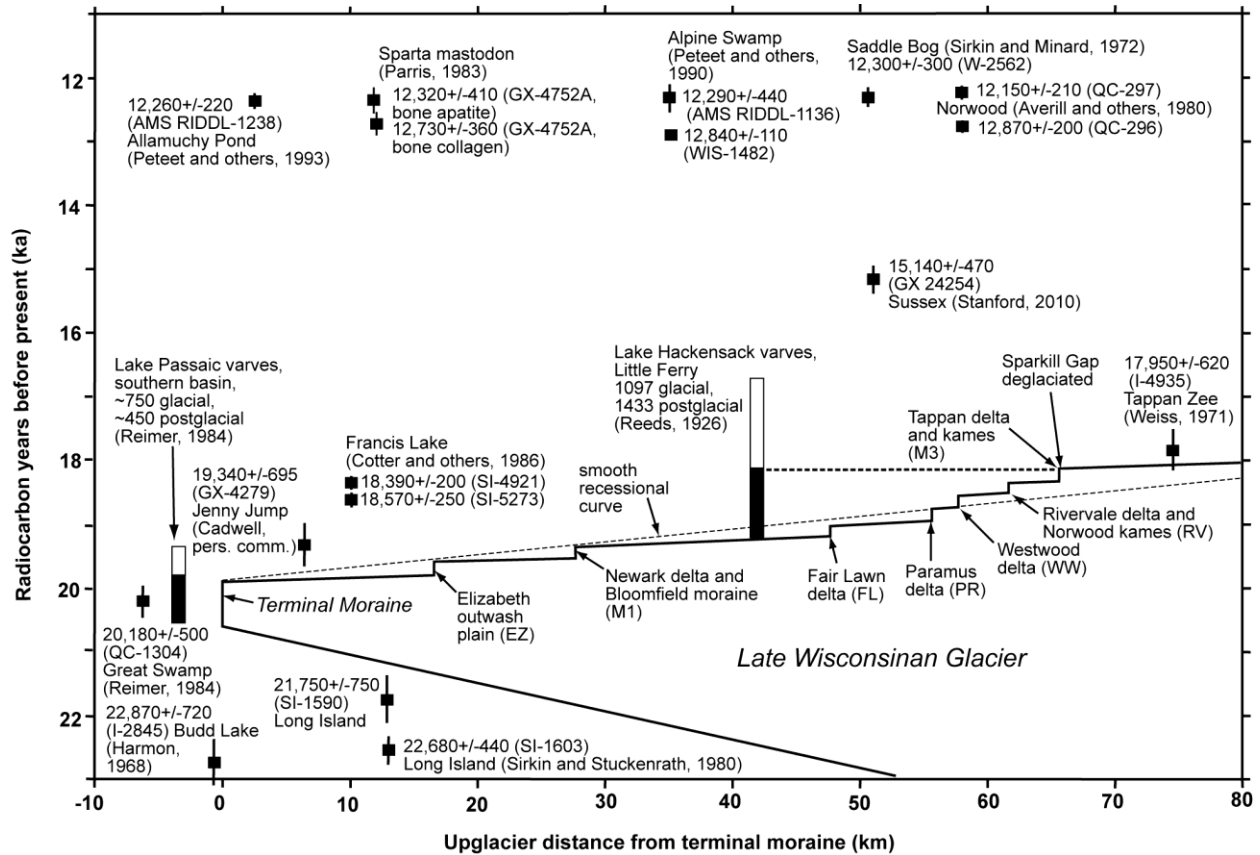

Figure 4. Time-distance plot of late Wisconsinan glaciation of study area. Radiocarbon dates are from the New Jersey-Long Island area. Recessional positions and ages are for the Hackensack lobe. Vertical lines at dates are one-sigma errors. Bold dotted line connecting top of glacial varves at Little Ferry and Sparkill dates the deglaciation of the gap and lowering of Lake Hackensack.

Lake Albany

The Hell Gate stage of Lake Albany extended up the Hudson valley to just north of the Hudson Highlands, where delta elevations (fig. 6) show the onset of an unstable phase of the lake. This unstable phase records lowering lake level, which indicates that the Narrows moraine dam was breached and an eroding spillway on till in the Narrows had replaced Hell Gate as the spillway. Rapid drainage of water from the Augusta stage of Lake Wallkill (fig. 5) at about 15.5 ka is the cause of the Narrows dam breach. The Augusta stage filled the ice-dammed north-draining Wallkill valley and was controlled by a spillway on the Wallkill-Delaware divide that drained southward into the Delaware basin (Stone and others, 2002). When the Moodna Creek valley at the north end of Schunnemunk Mountain, and Storm King in the main Hudson valley to the east, was deglaciated, the Augusta stage dropped 230 feet (70 m), releasing 6 mi³ (25 km³) of water into Lake Albany (Stanford, 2010). At this time Lake Albany covered an area of 79 mi² (205 km²), about the same as the present-day Hudson River downstream from Moodna Creek, and the sudden inflow would have temporarily raised the level of Lake Albany by 390 feet (120 m) if added all at

once. Topography of the Moodna Creek and Wallkill basins upstream of the outburst point suggests that perhaps 100 feet (30 m) of the 230 foot (70 m) fall occurred suddenly, which would have raised Lake Albany by 160 feet (49 m). Since the Hell Gate stage water level projects to the Narrows dam at -70 feet (-21 m) and the height of the moraine in the Narrows was around 50 feet (15 m), judging from the headlands bordering the Narrows today, the 120 feet (36 m) of freeboard on the dam would have been overtopped by the 160 foot (49 m) rise. Alternatively, if the outburst was not sudden, the added
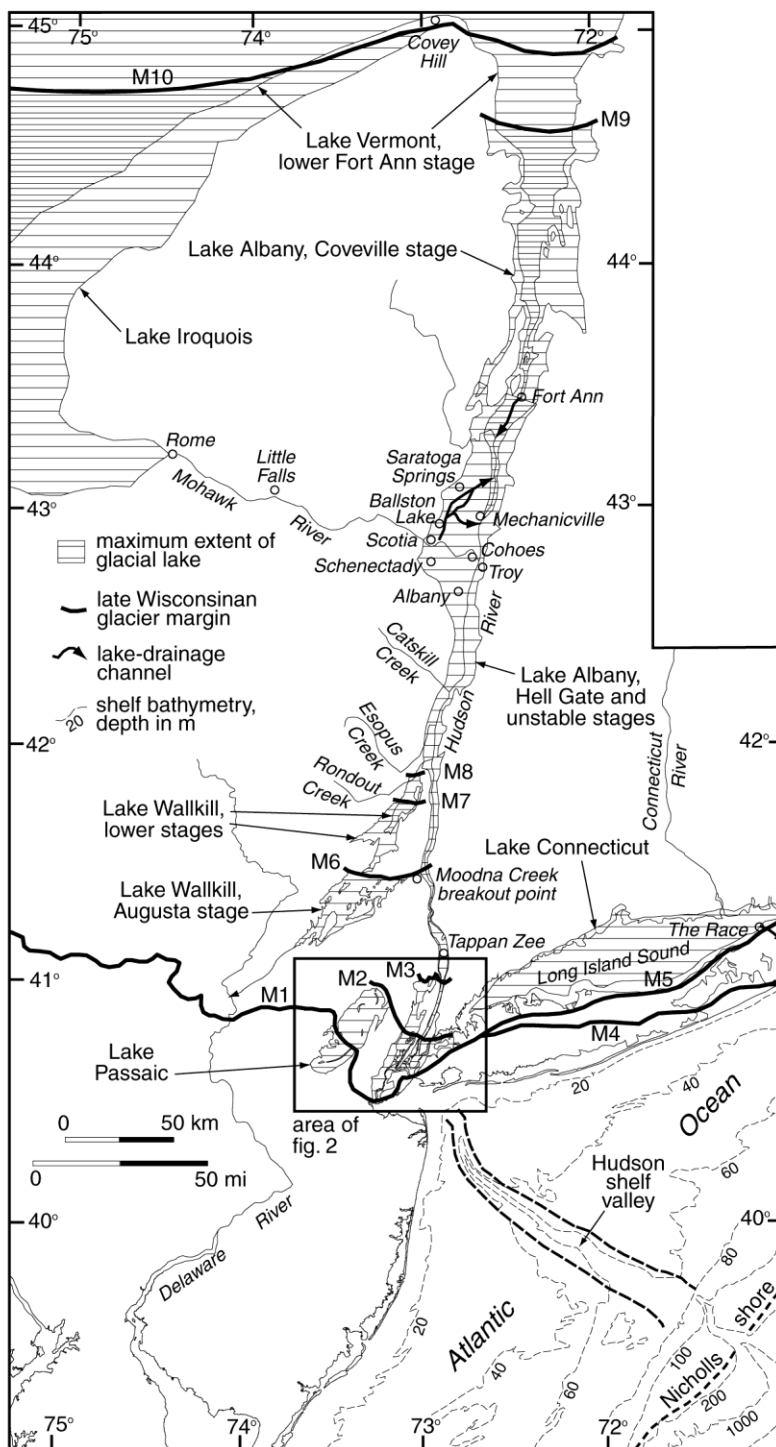
Figure 5. Lakes Wallkill, Albany, Iroquois, Vermont, and Connecticut, and the Hudson shelf valley. Ice margin abbreviations are: M1=terminal moraine in New Jersey and Pennsylvania, M2= last ice margin before Hell Gate

stage of Lake Albany was established, M3=last ice margin before Lake Hackensack lowers to Lake Albany, M4=Ronkonkoma Moraine, the late Wisconsinan terminus on Long Island, M5=Harbor Hill Moraine, M6=last ice margin before the Moodna Creek breakout flood, M7 and M8=last ice margins before draining of two lower stages of Lake Wallkill, M9=approximate last margin before stable Coveville stage of Lake Albany begins to lower, M10=last margin before Lake Iroquois lowers to Lake Vermont.

volume in Lake Albany could have caused seepage failure of the dam. If Storm King rather than Schunnemunk Mountain was the site of the outburst, which may have been the case depending on local ice-margin geometry, then there would be a slight increase in the released volume due to extension of Lake Wallkill into the Woodbury Creek valley between Schunnemunk Mountain and Storm King.
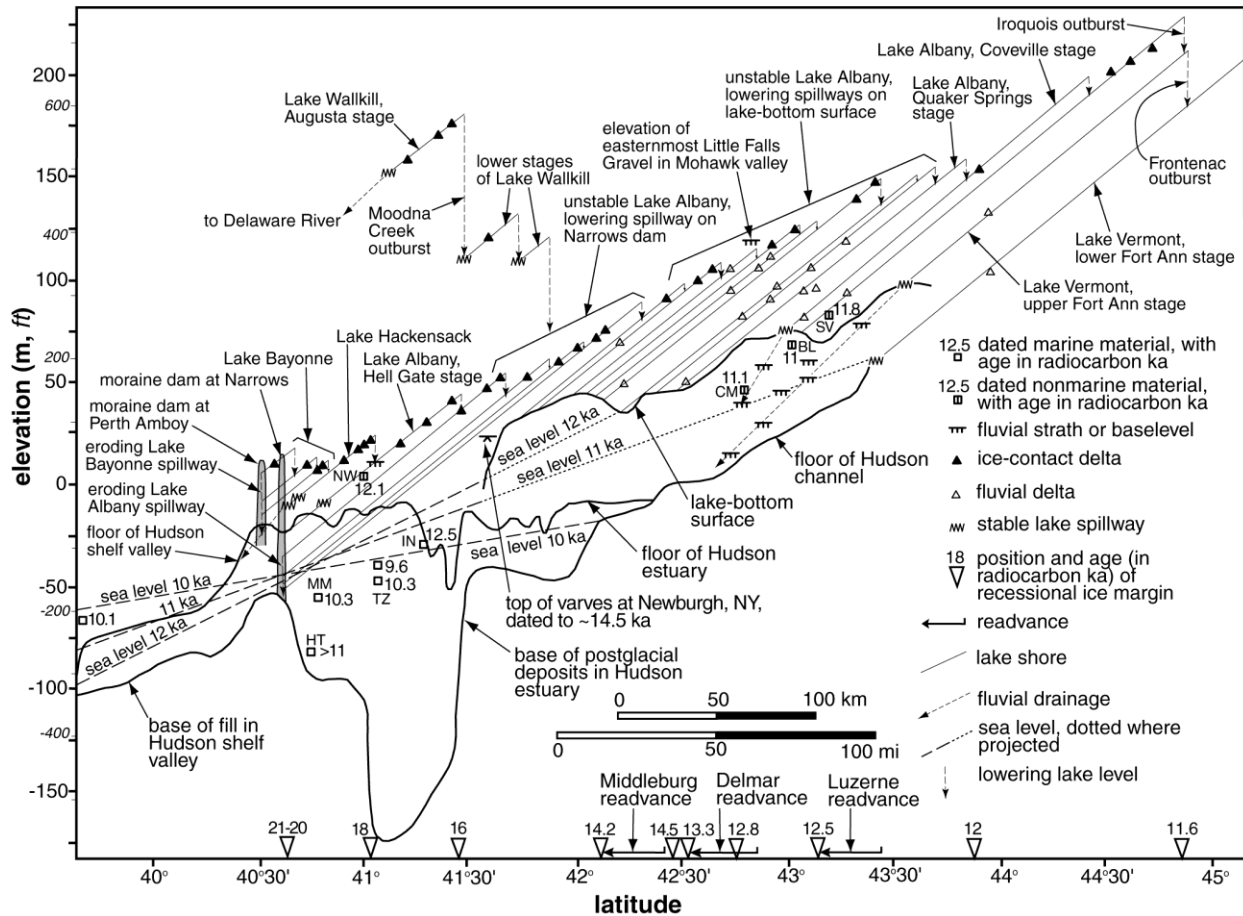
Figure 6. Present elevation of glacial-lake shorelines, lake-bottom surface of glacial Lake Albany, floor of the Hudson channel cut into the lake-bottom surface, top and bottom of postglacial fill in the Hudson estuary, and top and bottom of fill in the Hudson shelf valley. From Stanford (2010). Position and age of recessional ice margins and radiocarbon dates that bracket marine incursion are also shown. All elevations are projected to a north-south line perpendicular to isobases. Deltas, spillways, and fluvial features are compiled from: Woodworth (1905), Fairchild (1919), LaFleur (1965), Connally (1973), Connally and Sirkin (1970, 1973, 1986), Dineen and Hanson (1983), DeSimone and LaFleur (1986), Dineen (1986), Cadwell and Dineen (1987), Cadwell (1989), Stanford and Harper (1991), Wall and LaFleur (1994), Ridge (1997) and DeSimone and others (2008). Shorelines for the Coveville and Fort Ann stages north of 43°40' are from Rayburn and others (2005). Dates of recessional ice margins are from Ridge (2004, 2008), with minor additions in the lower valley. Base of postglacial fill in the estuary is from Worzel and Drake (1959), Newman and others (1969), Weiss (1974), and Lovegreen (1974). Base of fill in the shelf valley is from Thieler and others (2007). Radiocarbon dates are from Newman and others (1969), Averill and others (1980), Weiss (1974), Toney and others (2003), Donnelly and others (2005), Dineen and Miller (2006), and Miller (2008). Abbreviations for locations of dates are: NW=Norwood, NJ, MM=midtown Manhattan, HT=Holland Tunnel, TZ=Tappan Zee, IN=Iona Island, CM=Cohoes mastodon, BL=Ballstone Lake, SV=Schuylerville.

With the Narrows dam breached, Hudson valley drainage could now exit directly to the continental shelf in the New York Bight rather than around the east end of Long Island. This drainage cut the Hudson shelf valley, which leads from the Narrows to the edge of the continental shelf (fig. 5). The shelf in this region was subaerially exposed until about 10 ka, and the shelf valley itself was not flooded by marine incursion until 12 ka, leaving about 3500 years for fluvial incision of the shelf valley by Lake Albany outflows after dam breaching (fig. 6).

Erosion of the Narrows dam and the shelf valley was accelerated when Great Lakes outflows entered the Hudson via the Mohawk valley. During the Erie Interstade between 14.5 and 14.2 ka the Mohawk valley was temporarily deglaciated and conducted outflows from the Ontario and, possibly, the Erie basin into the Hudson (Ridge, 1997). This period of Great Lakes outflow corresponds to final erosion of the Narrows dam, to an elevation of about -200 feet (-60 m), because the altitude of deltas deposited at this time in the Albany area shows that the threshold for Lake Albany had shifted northward onto the emerging lake bottom north of the Hudson Highlands (fig. 6). From 14.2 to 12.7 ka readvances blocked the Mohawk valley but from its final deglaciation at 12.7 ka to 11 ka the Mohawk again conducted Great Lakes outflows into the Hudson valley, this time from Lake Iroquois in the Ontario basin (Wall and LaFleur, 1994). The threshold for Lake Albany continued to incise into emerging lake-bottom sediments downvalley, and migrated northward as isostatic rebound progressively raised the lake bottom to the south. The outflow from the threshold cut the channel that now contains the Hudson River in the mid-Hudson valley.

The rising sea had entered the lower Hudson valley by 12 ka (Newman and others, 1969), and from 12 to 10 ka, when rebound had elevated the upper valley above marine influence, high relative sea level in the lower valley limited the depth of fluvial incision in the Hudson channel upvalley. The inability of lake outflows to incise into the lake bottom during this interval created quasi-stable thresholds for the Quaker Springs and Coveville stages of Lake Albany and the upper Fort Ann stage of Lake Vermont (fig. 6).

When the north end of the Adirondack upland at Covey Hill was deglaciated at 10.9 ka, Lake Iroquois lowered in two stages (the "Iroquois outburst" and the "Frontenac outburst" on fig. 6) to the lower Fort Ann stage of Lake Vermont (Rayburn and others, 2005). These were enormous releases of 170 mi³ (700 km³) and 600 mi³ (2500 km³), respectively (Rayburn and others, 2005), which were the largest, and last, floods to discharge down the Hudson valley. High relative sea level in the Hudson valley at this time, however, prevented deep fluvial scour. By 10.6 ka, retreat opened the St. Lawrence valley and glacial sedimentation and meltwater discharge in the Hudson valley ceased.

History of Lake Passaic

The arcuate trace of the west-dipping cuesta ridges of the Watchung Mountains, formed on resistant basalt flows, reflects the canoe-shaped geometry of the Watchung syncline. The Ramapo Fault on its west side brings Proterozoic gneiss against the syncline. Erosion of the soft shale and sandstone enclosed between the basalts and the gneiss excavated a 30 mile long by 10 mile (48 by 16 km) wide basin between Second Watchung Mountain and the Highland Front. This basin is punctured by only two sets of gaps: the paired gaps at Little Falls and Paterson and at Short Hills and Millburn. These gaps are inheritances from the Pliocene course of the Hudson (see discussion on preglacial fluvial drainage above). Pre-intermediate drainage (fig. 7A) of the basin was established in the early Pleistocene after the pre-Illinoian glaciation and diversion of the Pensauken-Hudson river system. It exited the basin through the Short Hills-Millburn pair of gaps.

Illinoian glacial deposits in and south of the Short Hills gap indicate that Illinoian ice sealed this gap and thereby created an Illinoian version of Lake Passaic, probably similar to the late Wisconsinan Moggy Hollow stage. Illinoian delta and lacustrine-fan sand and gravel, and lake-bottom silt and clay, occur beneath late Wisconsinan deposits in the central and southern sections of the basin ("overramp zone' on fig. 8), where they fill the pre-intermediate valleys and also extend over the low interfluves between the valleys. The Illinoian fill in the Short Hills gap rises to about 200 feet (60 m) in elevation (section AA', fig. 8), so during Illinoian retreat a lake was maintained at that level until the gaps at Little Falls and Paterson were uncovered. This lake was much shallower than the late Wisconsinan recessional stages and limited the vertical accretion of Illinoian valley fill sediments north of the Short Hills gap. Erosion of the Illinoian valley fill during the Sangamon interglacial and early and middle Wisconsinan was minimal, because basalt in the Little Falls gap at an elevation of 180 feet (55 m) established a high base level for the basin, limiting fluvial incision.
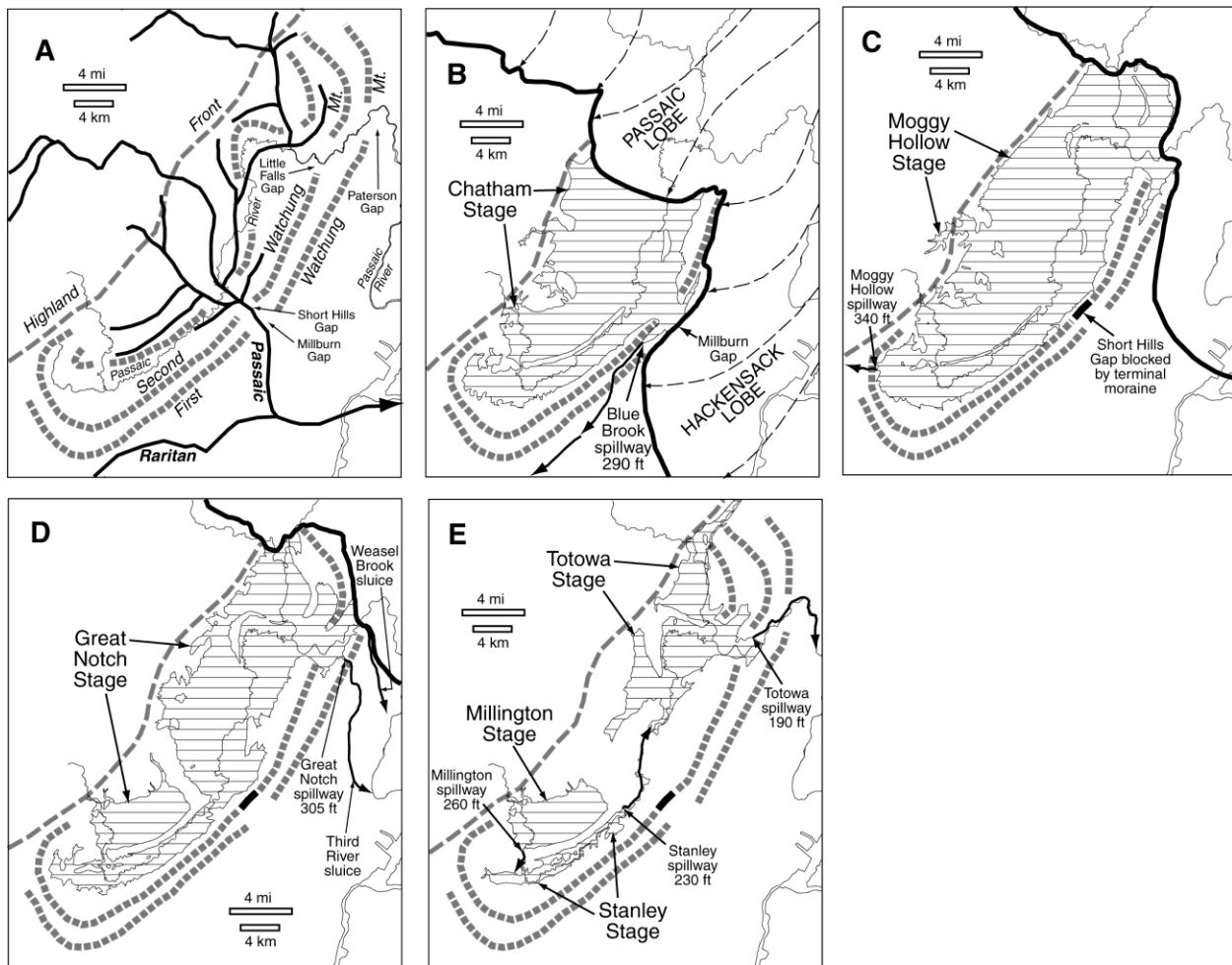
Figure 7. History of Lake Passaic. A. Fluvial drainage before the Illinoian glaciation. B. Advancing Hackensack lobe of late Wisconsinan glacier blocks Millburn gap, establishing the Chatham stage. Continued advance of Hackensack lobe blocks Short Hills gap, establishing the Moggy Hollow stage. C. Maximum extent of Moggy Hollow stage, just before uncovering of Great Notch. D. Maximum extent of Great Notch stage. Uncovering of Great Notch released 2.5 mi³ (10 km³) of water down the Third River sluice. Uncovering of Garrett Mountain released 1.2 mi³ (5 km³) of water down the Weasel Brook sluice. E. Postglacial stages.

The late Wisconsinan history of the lake includes three glacial lake stages (fig. 7B, C, D) and three postglacial lakes (fig. 7E). When the advancing Hackensack lobe, which extended further south than the Passaic lobe, blocked Millburn gap, the Chatham stage of the lake formed (fig. 7B), controlled by a

spillway at the head of the Blue Brook valley at an elevation of 290 feet (88 m) (Stop 3 of field trip). Filling of this lake stage buoyed the Passaic lobe ice and allowed the Passaic lobe to ramp over rather than erode pre-existing sediments as it advanced to the terminal position. These pre-existing sediments included Illinoian deposits and delta, fan, and lake-bottom sediments laid down in the Chatham stage in front of advancing ice (fig. 8, sections AA', BB'). An earlier, shallow, advance-phase lake probably formed when the Hackensack lobe blocked the Paterson gap. At this time the Passaic River was ponded to the level of the Illinoian fill in the Short Hills gap. However, since this fill was only about 20 feet (6 m) higher than the Little Falls base level, the lake was shallow, accumulated little sediment, and did not buoy the Passaic lobe.

Continued advance of the Hackensack lobe blocked the Blue Brook valley and then sealed the Short Hills gap. This caused the Chatham stage to rise 50 feet (15 m) to the Moggy Hollow stage (fig. 7C), which was controlled by a spillway into the Raritan basin across Second Watchung Mountain near Far Hills. The 50-foot (15 m) rise in lake level further buoyed the Passaic lobe, allowing it to ramp over the back end of a large fan-delta complex built into the lake at the terminal position. Till of the terminal moraine was deposited along the back end of this large delta deposit (fig. 8, sections AA', BB').

Deposition of till of the terminal moraine in the Short Hills gap filled the gap to an elevation of more than 400 feet (122 m), about 50 feet (15 m) higher than the Moggy Hollow stage, allowing this lake to expand northward as the ice front retreated. When Great Notch, a gap through First Watchung Mountain east of Little Falls, was uncovered, the Moggy Hollow stage dropped 80 feet (24 m) to the Great Notch stage (fig. 7D). This drop released about 2.5 mi³ (10 km³) of water down the Third River sluice downhill from Great Notch into the lower Passaic. The configuration of the glacier margin at this time (M1 on fig. 3, also on fig. 7C) is fixed by the last ice-contact deltas deposited in the Moggy Hollow stage: one along the Highland Front near Riverdale (Stop 4 of field trip) and one in the north end of the Preakness valley, where the ice front was lodged along the crest of Second Watchung Mountain.

Further retreat of the Hackensack lobe uncovered the north end of First Mountain at the Paterson gap (Stop 5 of field trip), allowing the Great Notch stage to drain down the Weasel Brook sluice. This flood released about 1.2 mi³ (5 km³) into the lower Passaic valley. Again, the position of the glacier margin at this time is fixed by Great Notch-stage deltas, including a delta at Riverdale (Stop 4 of field trip) and one at the north end of the Preakness valley. Both of these deltas are reworked from adjacent Moggy Hollow deltas by meltwater draining from local lakes adjacent to Lake Passaic, held in by ice deployed as shown in fig. 7D.

After the Great Notch stage drained, sediment dams held in three postglacial lakes. At Totowa, a large lacustrine fan blocked the Passaic valley in a narrow reach downstream from Little Falls, forming the dam for the Totowa stage. This lake extended up the Pompton and Ramapo valleys, where several large ice-contact deltas were deposited in it, including the Pompton plain, which nearly filled the northern bay of the Totowa lake, and a large delta which fills the Ramapo valley further north. Sand terraces were deposited by the Rockaway, Whippany, and Passaic rivers in the western and southern bays of the lake.

The terminal moraine formed a dam across the Passaic at Stanley, forming the Stanley stage. The Passaic and Dead rivers deposited silt and fine sand terraces in this shallow lake. The Great Swamp basin north of Lake Stanley was dammed by the terminal moraine-delta complex to the northeast. Postglacial Lake Millington filled this basin and was controlled by a spillway at Millington in a gap on Long Hill, the basalt ridge that forms the southern rim of the basin. Sand, reworked from the deltas to the north, was deposited in terraces in the northeast end of Lake Millington by Loantaka and Great Brooks.

Each of the postglacial lakes was controlled by spillways on erodible material (sand and gravel for Totowa, till for Stanley, and weathered and fractured basalt for Millington). The spillways were gradually lowered by erosion and the lakes eventually drained to leave the broad floodplains and marshes

that now occupy their floors. Today, these floodplains and marshes are valuable open spaces that provide thousands of acres of flood storage and wildlife habitat in an otherwise fully built environment.

Hydrogeology of the Lake Passaic Basin

The valley-fill deposits of the Lake Passaic basin produce over 15,000 million gallons per year and are the most productive glacial aquifers in New Jersey (Hoffman and Quinlan, 1994). They are the principal or sole water source for several municipal systems and so are classified by the EPA as a sole-source aquifer. The "overramp zone" (fig. 8), where Illionian and Chatham-stage lacustrine sand and gravel are thick, extensive, and fill the pre-intermediate buried fluvial valleys, is the most productive sector of the aquifer system (see distribution of production wells on fig. 8). The Illinoian and Chatham-stage sands and gravels are overlain by low-permeability silt and clay lake-bottom deposits and till. Till matrix is chiefly silty sand to sandy silt but is of low permeability because it has been highly consolidated in most places by the weight of overlying ice. These materials act as confining or semi-confining layers. Some wells bored through these sediments into underlying sand and gravel flowed at the surface when first drilled, indicating artesian conditions, although pumping in recent decades has greatly lowered the piezometric surface and wells no longer flow (Meisler, 1976; Hoffman and Quinlan, 1994). Gradients on the piezometric surface in the overramped sand and gravel indicate flow toward the Short Hills gap, along the original pre-intermediate fluvial valley gradient, although there is much perturbation of the natural gradient by pumping-induced cones of depression. Recharge to the confined, overramped sand and gravel is by infiltration of precipitation on the delta fronting the terminal moraine, which connects to the overramped sand and gravel in the subsurface (fig. 8, section BB'), by stream loss from the Passaic and Whippany rivers where they flow across the delta-moraine complex, and from vertical leakage of surface water through the lake-bottom sediment and till capping the overramped sand and gravel at and north of the moraine.

Neither Illinoian nor late Wisconsinan ice advanced south of the moraine complex (fig. 8, late Wisconsinan limit), and pre-Illinoian deposits in this area are thin, patchy erosional remnants largely above the level of the valley fill. Thus, no sand and gravel occurs in the valley fills south of the moraine complex. These fills instead consist of late Wisconsinan lake clays and silts deposited atop Illinoian lake clays and silts. These deposits confine the underlying bedrock but provide no water themselves.

North of the overramp zone, Passaic lobe ice was not buoyed by Chatham-stage lake water, since the Millburn gap was not yet blocked by Hackensack ice. The Passaic lobe thus eroded rather than overran any Illinoian stratified deposits on the landscape during advance. In fact, Passaic-lobe ice was so erosive in this sector that it scoured overdeepenings as much as 300 feet deep into shale bedrock in the lowlands between the basalt ridges (fig. 8, late Wisconsinan glacial overdeepenings). The valley fill in this northern sector of Lake Passaic thus consists exclusively of recessional lacustrine sediments on a basal till on bedrock, with no buried pre-advance materials (fig. 8, section BB', north end). The recessional lacustrine deposits are chiefly silt and clay lake-bottom sediments but in places, particularly in the valley fill along the Highland Front, permeable lacustrine-fan sand and gravel underlies the lake-bottom material. These fan deposits supply several municipal well fields in this area (Stop 4 of field trip). The fan gravels were deposited at the mouths of subglacial tunnels as the ice front retreated, and were later buried by lake-bottom sediment that accumulated vertically as fine sediment settled out of the lake. In places the fans rise above the lake-bottom fill and crop out as knolls and ridges (Stop 4 of field trip). In places along the Highland Front they built up far enough to reach the lake surface, where they prograded into the lake to form deltas. Other deltas along the Highland Front were deposited where ice-lateral meltwater channels draining down the Front entered the lake. The outcropping deltas and fans act as recharge conduits for the buried fans, where the deposits are physically connected.

Shallow-water delta sands deposited in the postglacial lakes by meteoric or meltwater streams, especially in the Totowa stage, lie above the thick deposits of lake clay laid down in the Moggy Hollow and Great Notch stages (fig. 8, north end of BB'). These sands may be unconfined aquifers if sufficiently thick, particularly adjacent to rivers, which are in hydraulic connection with the sands. However, no large-capacity wells tap these deposits in the Lake Passaic basin.
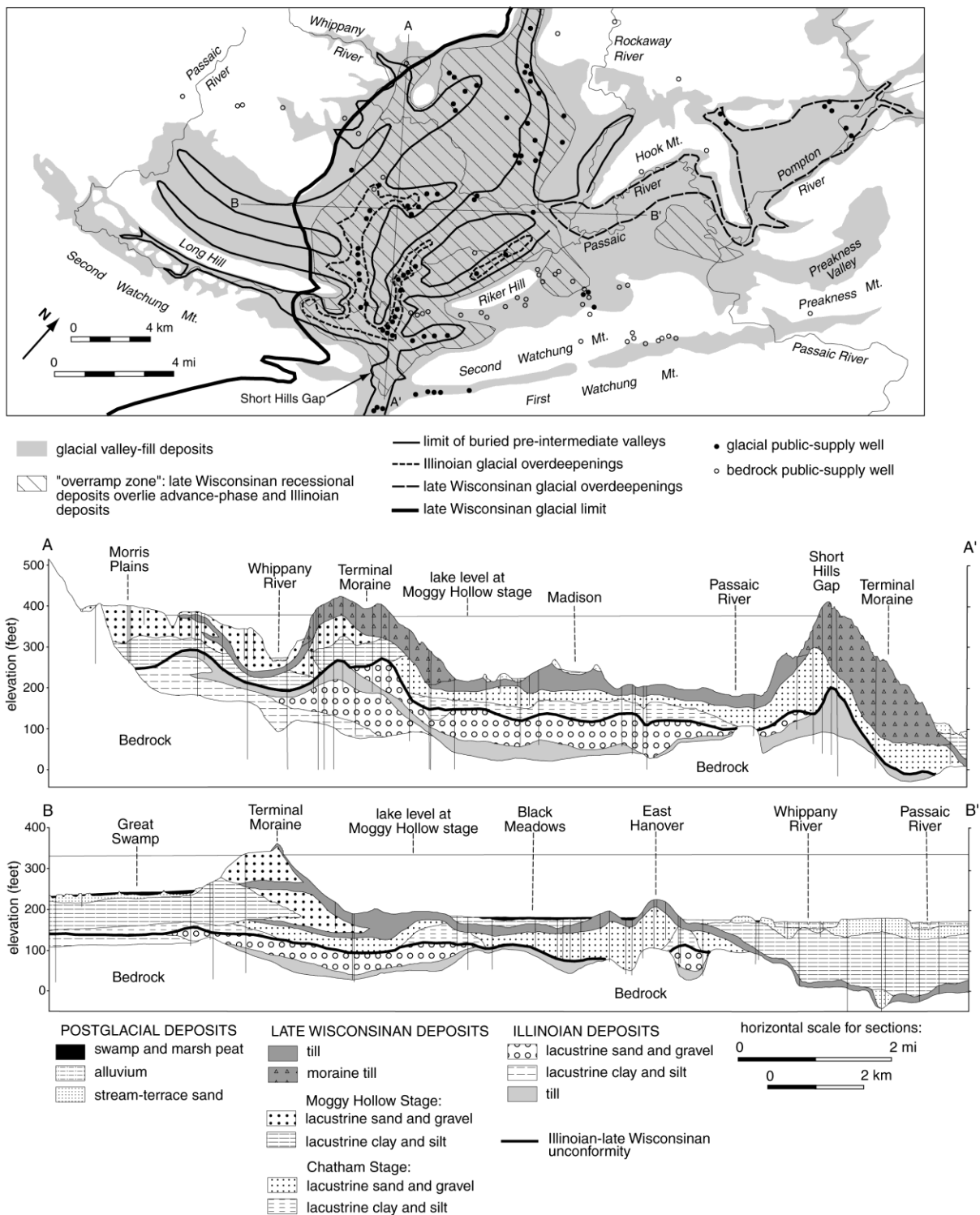

Figure 8. Stratigraphy and geomorphology of the valley-fill aquifer system in the Lake Passaic basin.

Summary

In the Pliocene the Hudson River, and a trunk river from southern New England, drained southwesterly from the New York City area to the Delmarva Peninsula. Drainage dislocations during the pre-Illinoian glaciation in the late Pliocene between 2 and 2.5 Ma redirected the Hudson seaward in the New York City area. Stabilization and incision of this drainage in the early Pleistocene established the lower Raritan and Passaic as tributaries to the Hudson, which exited to the shelf via a now-buried valley in Queens. An intermediate glaciation of probable Illinoian age (150 ka) reached the New York City area. Till and lacustrine and fluvial sediments in the lower parts of valley fills in the upper Passaic basin and on Long Island are deposits of this glacier. The late Wisconsinan glacier, centered as a southwesterly flowing lobe in the Hackensack lowland, arrived at its terminal position slightly after 21 ka and remained there for about 750 years, where it deposited a prominent moraine. It began to retreat from this position at 20 ka. Lake Passaic occupied the upper Passaic basin west of Second Watchung Mountain. It included an advance stage, a maximum stage, and a recessional stage. These stages spanned advance and retreat of the glacier. It drained in two stages eastward into Lake Bayonne. Stacked valley-fill sediments in Lake Passaic are important glacial aquifers. To the east, the terminal moraine formed a dam for glacial lakes during retreat. Lake Bayonne, the earliest lake, was controlled by an eroding spillway on the moraine dam at Perth Amboy. It was succeeded by Lake Hackensack and Lake Albany. Lake Hackensack drained eastward into Lake Albany when Sparkill Gap in the Palisades Ridge was deglaciated. Lake Albany was dammed by the moraine at the Narrows and extended up the Hudson valley as ice retreated. The Narrows dam was breached at 15.5 ka when Lake Wallkill flooded into the Hudson valley. Breaching initiated an unstable phase of the lake, with an eroding spillway first on the moraine, then on the emerging lake bottom in the mid-Hudson valley. Rising sea level flooded the lower Hudson valley after 12 ka, limiting fluvial incision and so providing quasi-stable spillways for the Quaker Springs and Coveville stages of Lake Albany, north of the Albany area, and the upper Fort Ann stage of Lake Vermont, in the Champlain lowland. Outburst floods from draining of Lake Iroquois at 10.9 ka discharged down the Hudson valley and were the last glacial events to affect the New York City area.

References Cited

Antevs, E., 1928, The last glaciation, with special reference to the ice sheet in North America: American Geographical Society, Research Series 17, 292 p.

Averill, S. P., Pardi, R. R., Newman, W. S., and Dineen, R. J., 1980, Late Wisconsinan-Holocene history of the lower Hudson region: new evidence from the Hackensack and Hudson river valleys: *in* Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: field guide of the 52nd annual meeting of the New York State Geological Association, p. 160-186.

Baskerville, C. A., 1994, Bedrock and engineering geologic maps of the New York County and parts of Kings and Queens counties, New York, and parts of Bergen and Hudson counties, New Jersey: U. S. Geological Survey Miscellaneous Investigations Series Map I-2306, scale 1:24,000.

Bayley, W. S., Salisbury, R. D., and Kummel, H. B., 1914, Description of the Raritan quadrangle, New Jersey: U. S. Geological Survey Geologic Atlas, folio 191, 72 p.

Berry, E. W., and Hawkins, A. C., 1935, Flora of the Pensauken Formation in New Jersey: Geological Society of America Bulletin, v. 46, p. 245-252.

Boellstorff, J., 1978, North American Pleistocene stages reconsidered in light of probable Pliocene-Pleistocene continental glaciation: Science, v. 202, p. 305-307.

Buxton, H. T., and Shernoff, P. K., 1999, Ground-water resources of Kings and Queens counties, Long Island, New York: U. S. Geological Survey Water-Supply Paper 2498, 113 p.

Cadwell, D. H., 1989, Surficial geology map of New York, lower Hudson sheet: New York State Museum Map and Chart Series 40, scale 1:250,000.

Cadwell, D. H., and Dineen, R. J., 1987, Surficial geology map of New York, Hudson-Mohawk sheet: New York State Museum Map and Chart Series 40, scale 1:250,000.

Connally, G. C., 1973, Surficial geology of the Glens Falls region, New York: New York State Museum Map and Chart Series 23, scale 1:48,000.

Connally, G. G., and Sirkin, L. A., 1970, Late glacial history of the upper Wallkill valley, New York: Geological Society of America Bulletin, v. 81, p. 3297-3306.

Connally, G. G., and Sirkin, L. A., 1973, Wisconsinan history of the Hudson-Champlain lobe, *in* Black, R. F., Goldthwait, R. P., and Willman, H. B., eds., The Wisconsinan Stage: Geological Society of America Memoir 136, p. 47-69.

Connally, G. G., and Sirkin, L. A., 1986, Woodfordian ice margins, recessional events, and pollen stratigraphy of the mid-Hudson valley, *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 50-72.

Cotter, J. F. P., Ridge, J. C., Evenson, E. B., Sevon, W. D., Sirkin, L. A., and Stuckenrath, R., 1986, The Wisconsinan history of the Great Valley, Pennsylvania and New Jersey and the age of the "Terminal Moraine", *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 22-50.

deLaguna, W., 1948, Geologic correlation of logs of wells in Kings County, New York: State of New York, Department of Conservation, Water Power and Control Commission, Bulletin GW 17, 35 p.

DeSimone, D. J., and LaFleur, R. G., 1986, Glaciolacustrine phases in the northern Hudson lowland and correlatives in western Vermont: Northeastern Geology, v. 8, p. 218-229.

DeSimone, D. J., Wall, G. R., Miller, N. G., Rayburn, J. A., and Kozlowski, A. L., 2008, Glacial geology of the northern Hudson through southern Champlain lowlands: Guidebook for the 71st annual reunion of the Northeastern Friends of the Pleistocene, 58 pp.

Dineen, R, J., 1986, Deglaciation of the Hudson valley between Hyde Park and Albany, New York, *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 89-108.

Dineen, R. J., and Hanson, E. L., 1983, Bedrock topography and glacial deposits of the Colonie channel between Saratoga Lake and Coeymans, New York: New York State Museum Map and Chart Series 37, 39 p.

Dineen, R. J., and Miller, N. G., 2006, Age and paleoecology of plant fossils associated with the Quaker Springs stage of Lake Albany, and a chronology of deglacial events in the Hudson-Champlain lowlands, New York: Geological Society of America Abstracts with Programs, v. 38, p. 8.

Donnelly, J. P., Driscoll, N. W., Uchupi, E., Keigwin, L. D., Schwab, W. C., Thieler, E. R., and Swift, S. A., 2005, Catastrophic meltwater discharge down the Hudson valley: a potential trigger for the Intra-Allerød cold period: Geology, v. 33, p. 89-92.

Fairchild, H. L., 1919, Pleistocene marine submergence of the Hudson, Champlain, and St. Lawrence valleys: New York State Museum Bulletin 209, 76 p.

Fuller, M. L., 1914, The geology of Long Island, New York: U. S. Geological Survey Professional Paper 82, 231 p.

Groot, J. J., and Jordan, R. R., 1999, The Pliocene and Quaternary deposits of Delaware: palynology, ages, and paleoenvironments: Delaware Geological Survey Report of Investigations 58, 36 p.

Harmon, K. P., 1968, Late Pleistocene forest succession in northern New Jersey: Ph.D. dissertation, Rutgers University, 203 p.

Hoffman, J. L., and Quinlan, J., 1994, Ground-water-withdrawal and water-level data for the central Passaic River basin: N. J. Geological Survey Geologic Report Series GSR 34, 78 p.

Johnson, D. W., 1931, Stream sculpture on the Atlantic slope: Columbia University Press, New York, 142 p.

LaFleur, R. G., 1965, Glacial geology of the Troy, N. Y., quadrangle: New York State Museum and Science Service, Map and Chart Series 7, 22 p.

Lewis, R. S., and Stone, J. R., 1991, Late Quaternary stratigraphy and depositional history of the Long Island Sound basin: Connecticut and New York: Journal of Coastal Research Special Issue 11, p. 1-23.

Lovegreen, J. R., 1974, Paleodrainage history of the Hudson estuary: M.S. thesis, Columbia University, 152 p.

Martino, R. L., 1981, The sedimentology of the late Tertiary Bridgeton and Pensauken formations in southern New Jersey: Ph.D. dissertation, Rutgers University, 299 p.

Meisler, H., 1976, Computer simulation model of the Pleistocene valley-fill aquifer in southwestern Essex and southeastern Morris counties, New Jersey: U. S. Geological Survey Water Resources Investigations 76-25, 76 p.

Merrill, F. J. H., Darton, N. H., Hollick, A., Salisbury, R. D., Dodge, R. E., Willis, B., and Pressy, H. A., 1902, Description of the New York City quadrangle: U. S. Geological Survey Geologic Atlas, folio 83, 19 p.

Miller, N. G., 2008, The Cohoes mastodon and Younger Dryas in eastern New York, in DeSimone, D. J., Wall, G. R., Miller, N. G., Rayburn, J. A. & Kozlowski, A. L., eds., Glacial geology of the northern Hudson through southern Champlain lowlands: guidebook for the 71st annual reunion of the Northeastern Friends of the Pleistocene, p. 19-25.

Newman, W. S., Thurber, D. H., Zeiss, H. S., Rokach, A., and Musich, L., 1969, Late Quaternary geology of the Hudson River estuary: a preliminary report: Transactions of the New York Academy of Sciences, v. 31, p. 548-570.

Obbink, E. A., Carlson, A. E., and Klinkhammer, G. P., 2010, Eastern North American freshwater discharge during the Bølling-Allerød warm periods: Geology, v. 32, p. 171-174.

Oldale, R. N., Cronin, T. M., Valentine, P. C., Spiker, E. C., Blackwelder, B. W., Belknap, D. F., Wehmiller, J. F., and Szabo, B. J., 1982, The stratigraphy, absolute age, and paleontology of the upper Pleistocene deposits at Sankaty Head, Nantucket Island, Massachusetts: Geology, v. 10, p. 246-252.

Owens, J. P., and Minard, J. P., 1979, Upper Cenozoic sediments of the lower Delaware valley and northern Delmarva Peninsula, New Jersey, Pennsylvania, Delaware, and Maryland: U. S. Geological Survey Professional Paper 1067D, 47 p.

Parris, D. C., 1983, New and revised records of Pleistocene mammals of New Jersey: The Mosasaur, v. 1, p. 1-21.

Perlmutter, N. M., and Geraghty, J. J., 1963, Geology and ground-water conditions in southern Nassau and southeastern Queens counties, Long Island, N.Y.: U. S. Geological Survey Water Supply Paper 1613A, 205 p.

Peteet, D. M., Vogel, J. S., Nelson, D. E., Southon, J. R., Nickmann, R. J., and Heusser, L. E., 1990, Younger Dryas climate reversal in northeastern USA? AMS ages for an old problem: Quaternary Research, v. 33, p. 219-230.

Peteet, D. M., Daniels, R. A., Heusser, L. E., Vogel, J. S., Southon, J. R., and Nelson, D. E., 1993, Late-glacial pollen, macrofossils, and fish remains in northeastern USA—the Younger Dryas oscillation: Quaternary Science Reviews, v. 12, p. 597-612.

Rayburn, J. A., Knuepfer, P. L. K., and Franzi, D. A., 2005, A series of large, late Wisconsinan meltwater floods through the Champlain and Hudson valleys, New York state, USA: Quaternary Science Reviews, v. 24, p. 2410-2419.

Rayburn, J. A., Franzi, D. A., and Knuepfer, P. L. K., 2007, Evidence from the Lake Champlain valley for a later onset of the Champlain Sea and implications for late glacial meltwater routing to the North Atlantic: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 246, p. 62-74.

Reeds, C. A., 1926, The varved clays at Little Ferry, New Jersey: American Museum Novitates, no. 209, p. 1-16. .

Reimer, G. E., 1984, The sedimentology and stratigraphy of the southern basin of glacial Lake Passaic, New Jersey: M. S. thesis, Rutgers University, 205 p.

Ridge, J. C., 1997, Shed Brook discontinuity and Little Falls Gravel: evidence for the Erie Interstade in central New York: Geological Society of America Bulletin, v. 109, p. 652-665.

Ridge, J. C., 2004, The Quaternary glaciation of western New England with correlations to surrounding areas, *in* Ehlers, J. & Gibbard, P. L., eds., Quaternary glaciations—extent and chronology, Part II: Elsevier, Amsterdam, p. 169-199.

Ridge, J. C. 2008: The North American glacial varve project (http://ase.tufts.edu/geology/varves), sponsored by the National Science Foundation and the geology department of Tufts University. Calibrated New England varve timescale as of April 2008.

Ridge, J. C., Braun, D. D., and Evenson, E. B., 1990, Does Altonian drift exist in Pennsylvania and New Jersey?: Quaternary Research, v. 33, p. 253-258.

Roy, M., Clark, P. U., Barendregt, R. W., Glasmann, J. R., and Enkin, R. J., 2004, Glacial stratigraphy and paleomagnetism of late Cenozoic deposits of the north-central United States: Geological Society of America Bulletin, v. 116, p. 30-41.

Salisbury, R. D., 1895, Surface geology: report of progress: N. J. Geological Survey Annual Report for 1894, p. 1-149.

Salisbury, R. D., 1902, The glacial geology of New Jersey: N. J. Geological Survey Final Report, v. 5, 802 p.

Salisbury, R. D., and Kummel, H. B., 1895, Lake Passaic: an exinct glacial lake: Journal of Geology, v. 3, p. 533-560.

Salisbury, R. D., and Peet, C. E., 1895, Drift phenomena of the Palisades ridge: N. J. Geological Survey Annual Report for 1894, p. 157-224.

Sasowsky, I. D., 1994, Paleomagnetism of glacial sediments from three locations in eastern Pennsylvania, *in* Braun, D. D., ed., Late Wisconsinan to pre-Illinoian (G?) glacial and periglacial events in eastern Pennsylvania: guidebook for the 57th Friends of the Pleistocene field conference, p. 21-23.

Schwab, W. C., Denny, J. F., Foster, D. S., Lotto, L. L., Allison, M. A., Uchupi, E., Danforth, W. W., Swift, B. A., Thieler, E. R., and Butman, B., 2003, High-resolution Quaternary seismic stratigraphy of the New York Bight continental shelf: U. S. Geological Survey Open-File Report 02-152, http://pubs.usgs.gov/of/2002/of02-152/.

Sirkin, L., 1986, Pleistocene stratigraphy of Long Island, New York, *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York. New York State Museum Bulletin 455, 6-21.

Sirkin, L. A., and Minard, J. P., 1972, Late Pleistocene glaciation and pollen stratigraphy in northwestern New Jersey: U. S. Geological Survey Professional Paper 800D, p. D51-D56.

Sirkin, L. A., and Stuckenrath, R., 1980, The Portwashingtonian warm interval in the northern Atlantic coastal plain: Geological Society of America Bulletin, v. 91, p. 332-336.

Soren, J., 1978, Subsurface geology and paleogeography of Queens County, Long Island, New York: U. S. Geological Survey Water Resources Investigations 77-34, 17 p.

Stanford, S. D., 1991, Surficial geology of the Roselle quadrangle, Union, Essex, and Morris counties, New Jersey: N. J. Geological Survey Open-File Map OFM 8, scale 1:24,000.

Stanford, S. D., 1993, Late Cenozoic surficial deposits and valley evolution of unglaciated northern New Jersey: Geomorphology, v. 7, p. 267-288.

Stanford, S. D., 2002, Surficial geology of the Elizabeth quadrangle, Essex, Hudson, and Union counties, New Jersey: N. J. Geological Survey Open-File Map OFM 42, scale 1:24,000.

Stanford, S. D., 2003, Surficial geology of the Paterson quadrangle, Passaic, Bergen, and Essex counties, New Jersey: N. J. Geological Survey Open-File Map OFM 54, scale 1:24,000.

Stanford, S. D., 2007a, Surficial geology of the Chatham quadrangle, Morris, Union, and Somerset counties, New Jersey: N. J. Geological Survey Open-File Map OFM 69, scale 1:24,000.

Stanford, S. D., 2007b, Surficial geology of the Pompton Plains quadrangle, Morris, Passaic, Essex, and Bergen counties, New Jersey: N. J. Geological Survey Open-File Map OFM 68, scale 1:24,000.

Stanford, S. D., 2008, Surficial geology of the Bernardsville quadrangle, Morris and Somerset counties, New Jersey: N. J. Geological Survey Open-File Map OFM 74, scale 1:24,000.

Stanford, S. D., 2010, Onshore record of Hudson River drainage to the continental shelf from the late Miocene through the late Wisconsinan deglaciation, USA: synthesis and revision: Boreas, v. 39, p. 1-17.

Stanford, S. D., and Harper, D. P., 1991, Glacial lakes of the lower Passaic, Hackensack and lower Hudson valleys, New Jersey and New York: Northeastern Geology, v. 13, p. 271-286.

Stanford, S. D., Ashley, G. M., and Brenner, G. J., 2001, Late Cenozoic fluvial stratigraphy of the New Jersey piedmont: a record of glacioeustasy, planation, and incision on a low-relief passive margin: Journal of Geology, v. 109, p. 265-276.

Stanford, S. D., Ashley, G. M., Russell, E. W. B., and Brenner, G. J., 2002, Rates and patterns of late Cenozoic denudation in the northernmost Atlantic coastal plain and piedmont: Geological Society of America Bulletin, v. 114, p. 1422-1437.

Stone, B. D., and Borns, H. W., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island, and adjacent Georges Bank and Gulf of Maine: Quaternary Science Reviews, v. 5, p. 39-53.

Stone, B. D., Stanford, S. D., and Witte, R. W., 2002, Surficial geologic map of northern New Jersey: U. S. Geological Survey Miscellaneous Investigations Series Map I-2540-C, scale 1:100,000.

Suter, R., deLaguna, W., and Perlmutter, N. M., 1949, Mapping of geologic formations and aquifers of Long Island, New York: New York State Water Power and Control Commission Bulletin GW-18, 212 p.

Thieler, E. R., Butman, B., Schwab, W. C., Allison, M. A., Driscoll, N. W., Donnelly, J. P., and Uchupi, E., 2007, A catastrophic meltwater flood event and the formation of the Hudson shelf valley: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 246, p. 120-136.

Toney, J. L., Rodbell, D. T., and Miller, N. G., 2003, Sedimentologic and palynologic records of the last deglaciation and Holocene from Ballston Lake, New York: Quaternary Research, v. 60, p. 189-199.

Volkert, R. A., 2010, Bedrock geologic map of the Pompton Plains quadrangle, Passaic, Morris, Bergen, and Essex counties, New Jersey: N. J. Geological Survey Geologic Map Series GMS 10-1, scale 1:24,000.

Wall, G. R., and LaFleur, R. G., 1994, Post-glacial paleochannel development in the Hudson/Mohawk lowlands, New York: Northeastern Geology, v. 16, p. 116-122.

Weiss, D., 1971, Late Pleistocene stratigraphy and paleoecology of the lower Hudson River estuary: Ph.D. dissertation, New York University, 160 p.

Weiss, D., 1974, Late Pleistocene stratigraphy and paleoecology of the lower Hudson River estuary: Geological Society of America Bulletin, v. 85, p. 1561-1570.

Woodworth, J. B., 1901, Pleistocene geology of a portion of Nassau County and Borough of Queens: N. Y. State Museum Bulletin 48, p. 617-668.

Woodworth, J. B., 1905, Ancient water levels of the Champlain and Hudson valleys: New York State Museum Bulletin 84, 265 p.

Worzel, J. L., and Drake, C. L., 1959, Structure section across the Hudson River at Nyack, New York, from seismic observations: Annals of the New York Academy of Sciences, v. 80, p. 1092-1105.

Road Log and Field Stops

0.0   0.0   Turn right on Victory Boulevard from College of Staten Island gate.

0.1   0.1   Turn right on South Gannon Ave. toward I-278 east.

0.4   0.3   Exit left onto I-278 east.

4.5   4.1   Exit right at Exit 15, last before Verrazano Bridge.

4.8   0.3   Left at light at end of ramp onto Lily Pond Ave. toward Bay Street. Lily Pond Ave. becomes School Road around bend.

5.3   0.5   Straight at light into parking lot for Arthur Von Briesen Park. STOP 1.

Stop 1. The Narrows

Walk up park paths over knoll and ridge topography of the terminal moraine to the viewpoint at the bluff top of the Narrows and Verrazano Bridge.

We are standing on the late Wisconsinan terminal moraine, a 1 to 2 mile-wide (1.6 to 3 km) belt of till between 50 and 150 feet (15 and 45 m) thick, with up to 100 feet (30 m) of ridge, knoll, and basin surface relief. Ice arrived at the moraine just after 21 ka (all dates stated in radiocarbon years), and varves counted in the southern basin of Lake Passaic indicate that the ice front stayed in the moraine belt for about 750 years (Reimer, 1984). The height of the headlands here and in Brooklyn indicates that the moraine was once continuous across the mile width of the Narrows at about 50 feet (15 m) in elevation. During late Wisconsinan retreat the Narrows dam and a moraine dam across the Arthur Kill at Perth Amboy, NJ held in glacial Lake Bayonne in the Arthur Kill-Kill van Kull-New York Bay lowland. Lake Bayonne was controlled by an eroding spillway on till at the Perth Amboy dam, and gradually lowered a total of about 50 feet (15 m) as ice retreated. When the retreating ice margin uncovered Hell Gate in the East River (M1 on fig. 3), Lake Bayonne drained eastward into the Long Island Sound lowland and Hell

Gate became the outlet for Lake Albany. The Hell Gate spillway is on gneiss at -30 feet (-9 m). The Hell Gate stage of Lake Albany was thus a stable lake and the shoreline, reconstructed from ice-contact deltas (text fig 6), projects to the Narrows at -70 feet (-21 m), indicating about 120 feet (36 m) of freeboard on the Narrows dam during the Hell Gate stage. The Hell Gate stage expanded northward up the Hudson valley to north of the Highlands, where delta elevations record the onset of an unstable phase of Lake Albany, with declining lake levels recording an eroding spillway (fig. 6). This indicates breaching of the Narrows dam and replacement of the stable Hell Gate spillway with an eroding spillway on moraine till.

Dam breaching was caused by an outburst flood at 15.5 ka from the Augusta stage of Lake Wallkill into Lake Albany. When the north end of Schunnemunk Mountain (and Storm King in the Hudson valley) was deglaciated, the Augusta stage lowered 230 feet (70 m), releasing 6 mi³ (25 km³) of water into Lake Albany. About 40% of this volume was probably added suddenly, producing a rapid rise in Lake Albany of about 160 feet (49 m), sufficient to overtop the Narrows dam.

Dam breaching also initiated fluvial erosion of the Hudson shelf valley, which leads from the Narrows to the edge of the continental shelf. The shelf valley was subaerially exposed until about 12 ka, providing 3500 years for fluvial incision of this prominent feature.

The moraine served as the spillway for Lake Albany until ice had retreated upvalley to just south of the Albany area, where delta elevations indicate that the spillway had shifted onto the emerging lake bottom in the Newburgh area, about 60 miles (100 km) up the valley from the Narrows. During its operation, the spillway on the moraine had eroded the dam to an elevation of -200 feet (-60 m). This depth allowed the rising sea to enter the Hudson valley at about 12 ka. From 12 to 10 ka high relative sea level in the lower Hudson valley limited fluvial incision of the lake bottom in the upper valley, providing quasi-stable spillways for the Quaker Springs and Coveville stage of Lake Albany, and the upper Fort Ann stage of Lake Vermont (fig. 6).

Major lake outflows from the Ontario basin discharged down the Mohawk valley and into the Hudson during the Erie Interstade between 14.5 and 14.2 ka, when the Mohawk valley was briefly deglaciated (Ridge, 1997), and again from 12.7 to 11 ka when the Mohawk valley was the outlet for Lake Iroquois (Wall and LaFleur, 1994). These outflows exited to the shelf through the Narrows and contributed to its erosion. When the Covey Hill threshold north of the Adirondacks was deglaciated, Lake Iroquois dropped in two stages to the lower Fort Ann level of Lake Vermont in the Champlain lowland, releasing 170 mi³ (700 km³) and 600 mi³ (2500 km³) of water at 10.9 ka (Rayburn and others, 2005). These megafloods also exited through the Narrows, and may have triggered the Younger Dryas stadial in the North Atlantic region between 10 and 11 ka. They were the last glacial events to affect the Hudson valley.

5.3	0.0	Proceed straight through light onto School Rd. from parking lot of Von Briesen Park.
5.5	0.2	Straight (on Lincoln Ave.) to I-278 west.
5.6	0.1	Straight on ramp onto I-278 west.
8.0	2.4	Serpentinite outcrop to left in roadcut.
11.2	3.2	Move right for 440 south.
11.4	0.2	Exit right to 440 south (exit 5).
13.8	2.4	Salt marsh along Arthur Kill visible to right. This is the floor of Lake Bayonne.

14.8	1.0	Enter Fresh Kills landfill, cells on both sides of 440. Former offload dock for garbage barges to right at bridge over Fresh Kills tidal creek. Landfill now closed and garbage is exported by truck and train.
16.6	1.8	Exit landfill area.
17.8	1.2	Climb onto an upland on Cretaceous deposits, veneered with till.
18.9	1.1	Exit right to stay on 440 south.
19.5	0.6	Exit right to Arthur Kill Road, exit 1.
19.8	0.3	Turn right at light at end of ramp onto Veterans Road.
20.0	0.2	Turn right at light onto Tyrellan Ave.
20.2	0.2	Turn right at light onto Boscombe Ave.
20.4	0.2	Proceed straight through light onto Page Ave.
21.0	0.6	Move to right lane so as to stay on Page through light.
21.1	0.1	Proceed straight through light on Page Ave.
21.6	0.5	Left at light onto Hylan Blvd.
22.3	0.7	Right into parking lot for Mount Loretto State Unique Area. STOP 2.

Stop 2. Red Bank Bluffs, Mount Loretto State Unique Area

NO SHOVELS, PICKS, OR HAMMERS

This stop requires a 1.25 mile walk along a beach and the possible fording of a shallow (ankle-deep) tidal creek. If you do not wish to do this, please remain on the bus, which will proceed 0.5 mile down Hylan Blvd. to meet us at Lemon Creek Park, at the end of the beach walk.

Walk south on driveway from parking area across the terminal moraine (fig. 9) to the beach on Raritan Bay, then walk northeastward along the beach to the parking lot for Lemon Creek Park, viewing bluff exposures along the way.

These bluffs provide an excellent exposure of the till that forms the terminal moraine. The bluffs were formed by wave erosion that has cut back about 1000 feet (300 m) into the outer edge of the moraine (fig. 9), which is estimated by fitting a smooth curve to the front of the moraine where it comes ashore to the north on Staten Island and to the west in New Jersey. The till exposed here is a compact, reddish-brown silty fine sand to silty fine-to-medium sand, containing 5-15% by volume pebbles and cobbles, and very few boulders. Pebble composition, counted on a total of 622 pebbles aggregated from five sites in the till along the bluff, is 54% red siltstone, 23% white to yellow well-rounded quartz pebbles, 13% gray siltstone and sandstone, 7% gneiss, 1% chert, and 1% purple conglomerate. The red siltstone is from the Passaic Formation in the Newark Basin. The quartz pebbles are from the Pensauken Formation, a

73

Pliocene fluvial gravel that was widespread in the New York City area before glaciation and is locally preserved in erosional remnants beneath till on Staten Island and in adjacent New Jersey. The gray siltstone and sandstone are from gray beds in the Newark Basin and from Paleozoic sedimentary rock in the Green Pond outlier and Wallkill valley. The gneiss is from Proterozoic bedrock in the Hudson Highlands. The chert is from Paleozoic carbonate rock in the Wallkill valley. The purple conglomerate is from the Green Pond and Skunnemunk formations in the Green Pond outlier, although some may be from Newark Basin fanglomerates. The reddish matrix color is derived from erosion of the red siltstone and sandstone of the Newark Basin. Notably absent or rare are basalt, diabase, and serpentinite, all local bedrock types found within 5 to 20 miles of the bluffs. Glacial flowlines, however, show that ice moving down the Hackensack lowland to this location did not cross those lithologies (fig. 2).

Also observed in the bluff (locations 2 and 4, fig. 9) are lenses or blocks of thinly bedded white and gray silt, fine sand, and coarse sand. The lenses at location 4 are within a deformed body of glaciofluvial sand and gravel, the one at location 2 appears to be within till, although it is situated behind an old sea wall and may have been artificially emplaced. These are glacially thrust blocks of Cretaceous sand and silt of the Magothy Formation, which, along with the underlying Raritan Formation, onlaps the bedrock in the southern half of Staten island and directly underlies the till. The undeformed bedding in the thrust blocks perhaps indicates that they were frozen when glacially entrained.
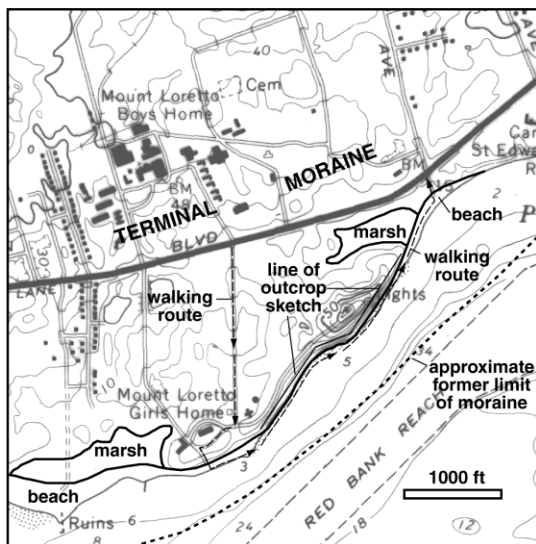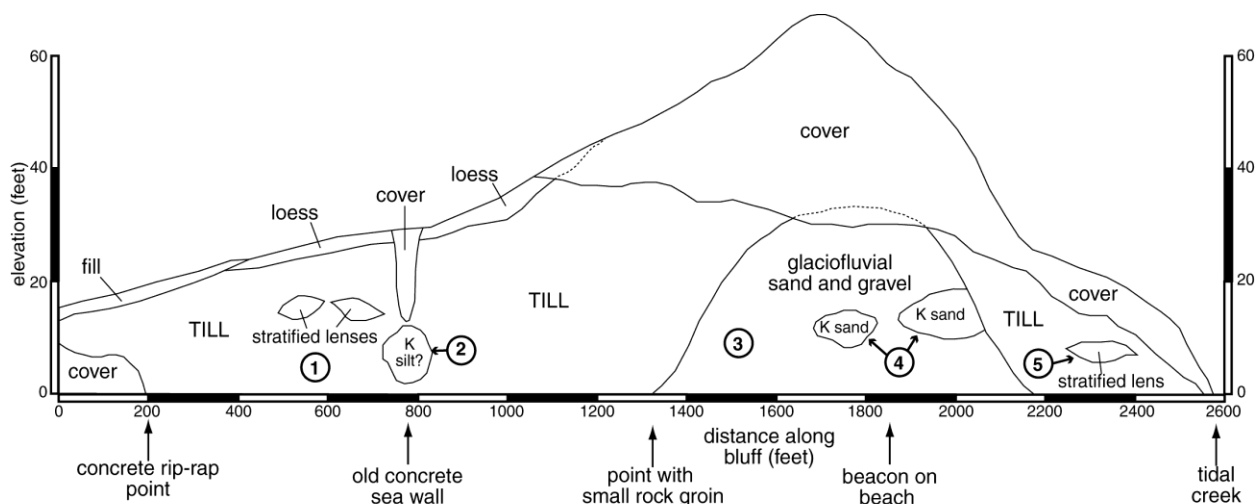
NOTEWORTHY OUTCROP FEATURES

(1) Good till exposures due to wave undercutting. Lenses of thinly bedded to laminated silt and sand within till.

(2) Gray, laminated silt and very fine sand. Clast of Cretaceous sediment within till or fill?

(3) Oversteepened dips and possible folding in cross-bedded glaciofluvial sand and gravel indicating glacial deformation.

(4) White to yellow, thin-bedded, micaceous fine sand, silt, and coarse sand. Glacially thrust Cretaceous deposits?

(5) Good till exposures with lenses of thinly bedded silt and sand.

Figure 9. Map and outcrop sketch of Red Bank bluffs, Mount Loretto State Unique area, Stop 2. Base map from USGS Arthur Kill 7.5-minute quadrangle.

The 70-foot hill is cored with light gray to pinkish-gray to very pale brown, cross bedded, clean sand and pebble-to-small cobble gravel, overlain by till. Dips on the cross-bed sets are oversteepened in places (location 3), and recumbent fold axes are also visible on the bluff face in some of the finer-grained beds, indicating that the sand and gravel has been deformed. This gravel is proglacial glaciofluvial outwash that was overrun and deformed by the advancing glacier. The deformation thickened the gravel here to form the core of the hill. Glaciofluvial plains front the moraine when it comes ashore further north on Staten Island, on Long Island, and at Perth Amboy, New Jersey. This deposit is of similar origin. Pebble composition in this unit, counted on a total of 366 pebbles aggregated from three sites in the exposure, is 41% red siltstone, 41% quartz pebbles, 13% gray siltstone and sandstone, 4% gneiss, and 1% purple conglomerate. The greater quartz-pebble content in the gravel compared to the till is from meltwater reworking of Pensauken Formation gravel that was likely widespread on the land surface in front of advancing ice.

Lenses of laminated silt and sand occur within the till at several places (locations 1 and 5). These lenses are about 20 to 50 feet long and generally less than 5 feet thick. They are approximately the same dimensions as the shallow basins on the moraine surface (visible in the meadows on the walk in) and may be swale-pond fills on the moraine surface that were later overrun by readvances as the ice front moved back and forth in the moraine belt. Again, lack of deformation suggests the fills were frozen when overrun.

22.3	0.0	Turn left onto Hylan Blvd. from parking lot.
23.0	0.7	Turn right at light onto Page Ave.
23.6	0.6	Move to right to stay on Page at light.
23.8	0.2	Cross Mill Creek, here a tidal creek. The head of this valley to the east (right) was an early spillway for Lake Bayonne, before it shifted to the moraine dam at Perth Amboy.
24.3	0.5	Turn left at light onto 440 north.
24.6	0.3	Bear right to 440 north.
32.1	7.5	Move left for I-278 west and Goethals Bridge.
32.3	0.2	Exit left for I-278 west and Goethals Bridge.
33.9	1.6	Cross the Arthur Kill. Railroad bridge to right is the Arthur Kill lift bridge, longest center-lift bridge in the world. It was built by the B&O railroad in 1959. Piers for it and the Goethals Bridge are on red shale at a depth of 30 to 40 feet.
34.4	0.5	Bayway refinery on left.
34.6	0.2	Exit left onto NJ Turnpike.
35.3	0.7	Bear left onto NJ Turnpike north after tolls.
38.4	3.1	Newark Airport to left, Port Newark to right. Both are built on former salt marsh atop

Lake Bayonne lake-bottom clays.

40.2	1.8	Exit right at Exit 14 for I-78 west.
40.7	0.5	Bear left for I-78 west.
41.6	0.9	Bear left onto I-78 west after toll.
41.9	0.3	Bear right to I-78 west.
43.2	1.3	Avoid 2-lane left exit for Clinton Ave., Newark. Rise from lake bottom onto a till on sandstone upland.
46.6	3.4	Cross Garden State Parkway.
50.1	3.5	Move left to stay on I-78 west.
50.6	0.5	Exit left to stay on I-78 west.
51.2	0.6	Roadcut in Orange Mountain basalt, of Lower Jurassic age. This basalt holds up First Watchung Mountain.
52.1	0.9	Roadcut through terminal moraine (behind soundwalls).
53.2	1.1	Cross Blue Brook spillway for Chatham stage of Lake Passaic, here buried by a small outwash plain laid down after the spillway became inactive.
53.5	0.3	Roadcut on right (to 55.6) in Preakness basalt, holding up Second Watchung Mountain.
55.6	2.1	Exit right onto Diamond Hill Road (exit 43).
56.2	0.6	Turn right at light onto McMane Ave.
57.0	0.8	Turn left at light onto 527 north (Glenside Ave.)
57.1	0.1	Proceed straight through light, staying on Glenside.
58.3	1.2	Turn right on 645 south (W. R. Tracy Drive).
58.7	0.4	Cross Chatham-stage sluiceway, occupied by a man-made lake here.
59.1	0.4	Turn right into Watchung Loop picnic area.
59.2	0.1	Comfort station and picnic site. STOP 3.

Stop 3. Watchung Reservation. Lunch and Pre-Illinoian Till.

NO SHOVELS OR PICKS

Before lunch at the picnic area, walk down trail towards Tracy Drive to the head of a gully with a rock-gabion floor, descend the gully to view exposures of till, then exit left at bottom of gully and return uphill on adjoining gravel trail to the picnic area.
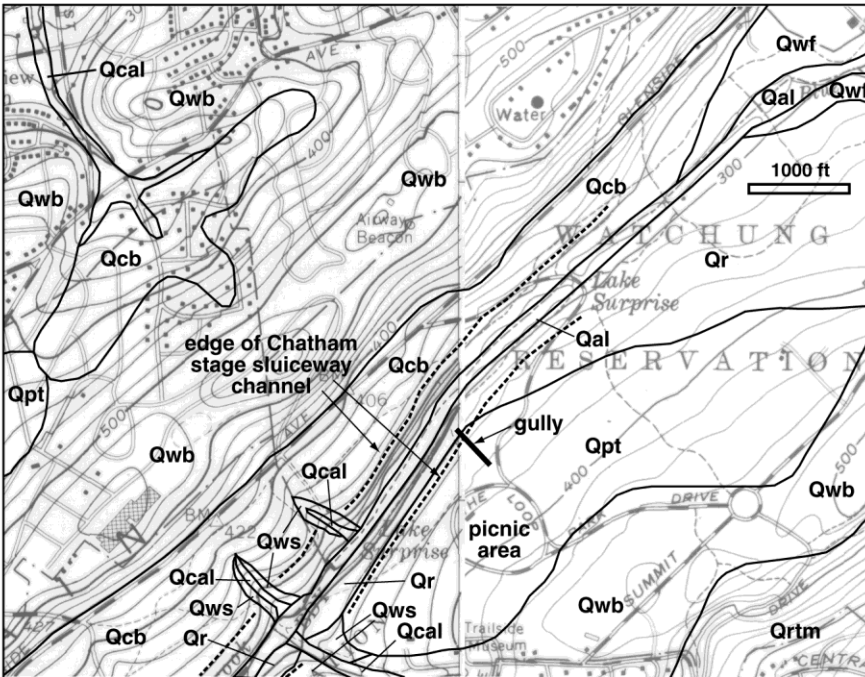
Figure 10. Map of area around Stop 3. Map units are: Qpt=pre-Illinoian till, Qr=late Wisconsinan Rahway Till, Qrtm=Rahway Till of the late Wisconsinan terminal moraine, Qwf=late Wisconsinan glaciofluvial sand and gravel, Qcb=basalt colluvium, Qal=postglacial alluvium, Qcal=alluvium and colluvium, undivided, Qwb=weathered basalt, Qws=weathered shale. Geology from Stanford (1991, 2007a). Base map from USGS Chatham and Roselle 7.5-minute quadrangles.

This gully is eroded into the easternmost deposit of pre-Illinoian till in New Jersey. This deposit is on the gentle dip slope of First Watchung Mountain, just outside the late Wisconsinan limit. Late Wisconsinan till (Qr on fig. 10), laid down by a brief advance beyond the terminal moraine, onlaps and overprints the pre-Illinoian till just to the north. The pre-Illinoian till here is a reddish-brown to reddish-yellow sandy clayey silt with deeply weathered to saprolitized pebbles and cobbles of red and gray siltstone, basalt, and gneiss. Gray quartzite, purple quartzite-conglomerate, black to brown chert, and well-rounded white to yellow quartz pebbles are unweathered. Large cobbles and boulders of gneiss are also intact, perhaps because the outermost weathered parts have spalled off. Fresh exposures observed in 1991 showed a thin tongue of late Wisconsinan till with lightly weathered to unweathered siltstone, gneiss, and basalt clasts overlying weathered micaceous siltstone of the Feltville Formation, at the downstream end of the gully. These materials are no longer well exposed here but can be seen in cuts along the valley side to the south. These observations show that the pre-Illinoian till rests on a gently sloping bedrock bench about 30 feet above the valley floor.

This landscape position is somewhat anomalous for pre-Illinoian till. Generally it is preserved only on flat surfaces on divides or hilltops. The anomaly is explained by the unusual history of this valley. The Blue Brook valley was deepened and widened around 21 ka because it served as the sluiceway (dotted line on fig. 10) for the Chatham stage of Lake Passaic (fig. 7). Before this incision, bedrock-surface elevation contours (Stanford, 1991) indicate that this was a broad divide in the intermontane valley between First and Second mountains, with a much smaller Blue Brook draining southwestward from the divide and another small stream draining northeastward in a valley now buried by the terminal moraine and a small fronting outwash plain. The divide area was protected from erosion, preserving the

pre-Illinoian till. The valley here is thus only 21,000 years old, not >800,000 years old like the other valleys in the pre-Illinoian glacial terrain. When the Hackensack lobe of the late Wisconsinan glacier blocked the Short Hills gap, the Blue Brook spillway shut down and the ice front here, no longer held at bay by the rushing lake outflow, briefly advanced into the now-empty sluice, depositing the thin tongue of till observed within the channel.

|		
59.2	0.0	Proceed to right around one-way loop road.
59.5	0.3	Bear left at end of loop onto 645 north (W. R. Tracy Drive).
60.4	0.9	Turn left at stop sign onto 527 south (Glenside Ave.)
61.7	1.3	Turn right at light onto I-78 east.
65.0	3.3	Move right into local lanes of I-78 east.
65.7	0.7	Move left to exit for NJ 24 west.
66.0	0.3	Exit left to NJ 24 west (exit 48).
67.2	1.2	Climb up terminal moraine plugging the former Short Hills gap.
67.8	0.6	Crest of moraine in Short Hills gap. Enter Lake Passaic basin.
69.4	1.6	Cross Passaic River, on floor of Lake Passaic.
71.7	2.3	Roadcuts from hereabouts to 72.2 are in lacustrine-fan deposits laid down in Moggy Hollow stage of Lake Passaic.
73.1	1.4	Black Meadows marsh to right, on floor of Lake Passaic. Terminal moraine forms rise in front. Training facility for NY [sic] Jets football team to left.
75.0	1.9	Roadcut through moraine. Till of the moraine hereabouts caps a section of stacked Illinoian and late Wisconsinan lacustrine sediment about 350 feet thick, one of the thickest glacial sections in NJ.
75.6	0.6	Exit right onto I-287 north.
79.7	4.1	Cross under I-80. Buried pre-intermediate Passaic-Rockaway valley here, filled with 200 feet of stacked late Wisconsinan and Illinoian till and lacustrine deposits.
82.3	2.6	Cross Rockaway River, redirected in postglacial time from the valley at 79.7.
83.6	1.3	Top of Boonton delta, deposited in Moggy Hollow stage of Lake Passaic.
87.5	3.9	Roadcut in another Moggy Hollow stage delta at Brook Valley Road overpass.
87.7	0.2	Outcrops of gneiss on left along Highland Front.
89.9	2.2	Roadcut in yet another Moggy Hollow stage delta, to be discussed at Stop 4.

90.5   0.6   Exit right to NJ 23 south (exit 52A).

91.2   0.7   Exit right for alternate 511 (Boulevard) and then another right onto West Parkway.

91.8   0.6   Turn right into parking lot for Foothills Park. STOP 4.

Stop 4. Lake Passaic Fan-Delta Complex, Foothill Park, Pequannock

Walk from parking lot to the southwest corner of the park lawn, near wall of former gravel pit, to the left of the striated gneiss outcrop. When returning to parking lot, visit the outcrop, which is interlayered white, buff, and light green quartz-plagioclase gneiss and dark gray amphibolite of Middle Proterozoic age. The gneiss and amphibolite show ductile deformation and mylonitic fabric because this outcrop is very close to the Ramapo Fault, the bounding normal fault on the west edge of the Newark Basin (Volkert, 2010).

Foothill Park and the office complex to the east over to NJ 23 are within a former gravel pit. This pit was dug into a lacustrine fan deposited in the Moggy Hollow stage of Lake Passaic. The fan formerly rose as much as 50 feet above the present land surface. The excavated part of the fan is the top of a deposit more than 200 feet thick (fig. 11, section AA'), extending down to a thin basal layer of till resting on bedrock. The pit exposures, and logs of test and production wells adjacent to and within the former pit, show that the fan is pebble-to-cobble gravel and sand with some interbeds of finer sediment, for its entire thickness.

Pebble composition of the fan gravel, counted on a total of 737 pebbles aggregated from 5 sites in the former pit, is 64% gneiss, 31% gray sandstone and siltstone, and 5% purple conglomerate. Gneiss is from the Highlands, adjacent to the deposit to the north and west. Gray sandstone and siltstone are from Paleozoic rock in the Wallkill valley and Green Pond outlier to the north. Purple conglomerate is from the Green Pond outlier. Notable is the absence of red siltstone and basalt from the Newark basin, which underlies most of the fan complex and the broad lowland to the east. Ice flow during glacial advance here, as recorded by striations in the vicinity (fig. 11), including those on the outcrop in the park, was southward from the Highlands into the Newark Basin. Most ice flow from the Highlands into the Newark Basin was blocked by earlier-arriving ice of the Passaic lobe flowing southwesterly in the Basin (fig. 2), so the pattern here is somewhat anomalous. We are near the mouth of the broad, deep, south-trending Wanaque River valley that cuts the Highlands just to the north. This valley helped channel Highlands ice out into the Newark Basin in this area. The Highlands ice deposited gneiss-rich till (Netcong Till, Qn on fig. 11) on the red siltstone bedrock hereabouts. This till also contains significant amounts of gray siltstone and purple conglomerate. Subglacial meltwater feeding the fan complex during deglaciation eroded this till, and gneiss bedrock just to the north in the Wanaque valley, rather than the local red siltstone, accounting for the Highlands provenance of the fan gravel.
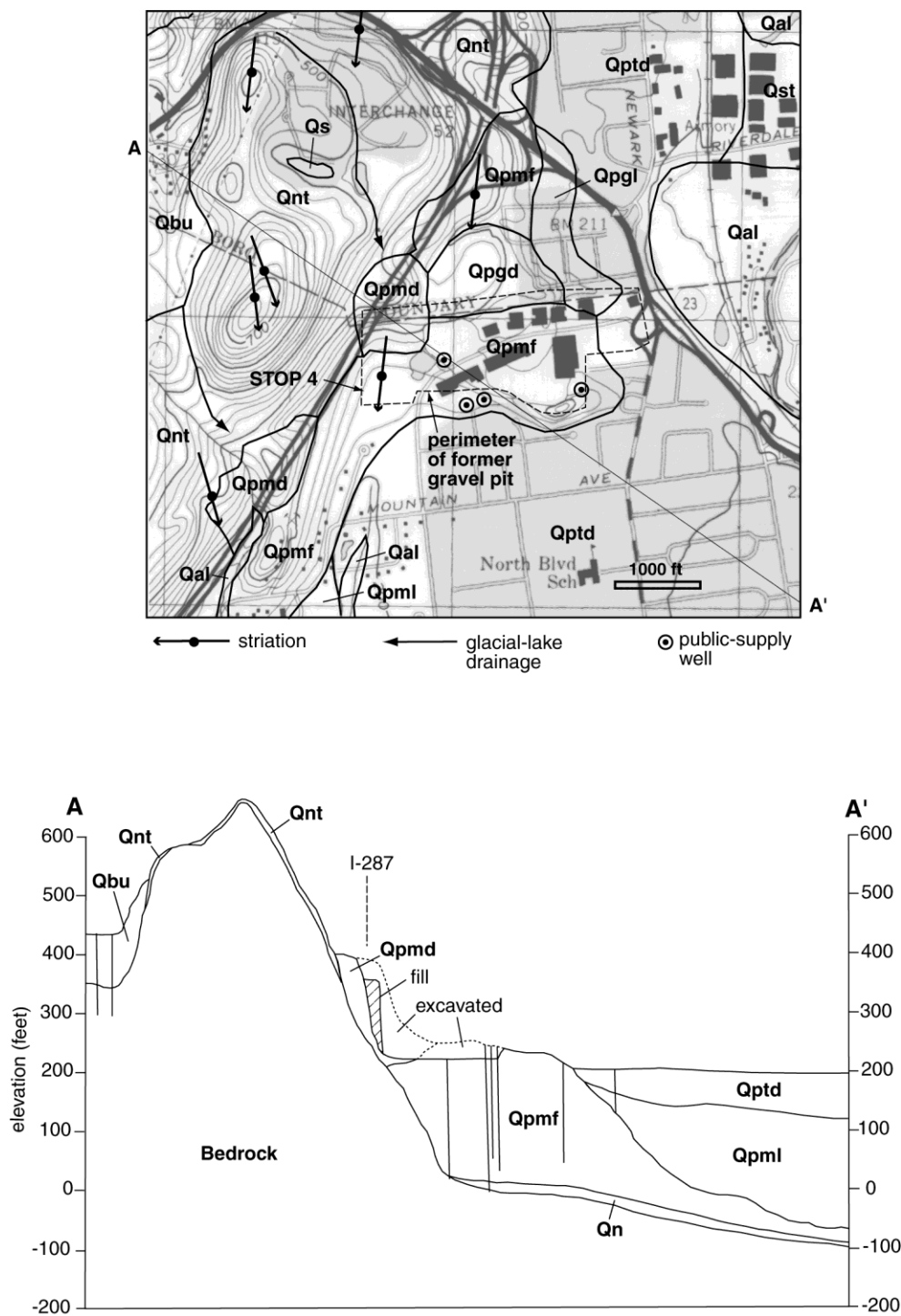
Figure 11. Map and section of area around Stop 4. Map units are: Qpmd=deltaic sand and gravel deposited in the Moggy Hollow stage, Qpmf=Lacustrine-fan sand and gravel deposited in the Moggy Hollow stage, Qpml=lake-bottom silt and clay deposited in the Moggy Hollow stage, Qpgd=deltaic sand and gravel deposited in the Great Notch stage, Qpgl=lake-bottom silt and clay deposited in the Great Notch stage, Qptd=deltaic sand and gravel deposited in the Totowa stage, Qbu=deltaic sand and gravel deposited in Lake Butler, Qnt=patchy Netcong Till and bedrock outcrop, Qn=Netcong Till, Qal=postglacial alluvium, Qst=postglacial stream-terrace sand. Geology from Stanford (2007b). Base map from USGS Pompton Plains 7.5 minute quadrangle.

The northwest edge of the pit was dug into a small Moggy Hollow stage delta. This delta is now largely gutted by the I-287 cut and fill (at mile 89.9 of road log) but a fragment west of I-287 marks its original surface at about 420 feet (128 m) in elevation, indicating that Lake Passaic here was about 500 feet (152 m) deep before deposition of recessional lacustrine deposits (fig. 11, section AA'). The north wall of the pit is dug into a Great Notch stage delta with a top surface at 320 feet (97 m), visible as the treeline on the hill across the park to the north. The pit wall here shows delta foreset sand overlying pebble-to-cobble fan gravel, indicating that the delta prograded over the older fan deposit. Fan deposits laid down in the Moggy Hollow stage also crop out north of the delta (fig. 11), indicating that this delta was not deposited in contact with the glacier. Rather, as the ice front retreated northward from the fan complex toward the present position of NJ 23, an outlet channel for Lake Butler, a local lake that occupied a north-draining valley just to the west of, and higher than, Lake Passaic, was opened at the 500-foot (152 m) elevation on the hill to the west of the fan complex (arrowed line on fig. 11). At the same time, Hackensack lobe ice on the east side of the Lake Passaic basin uncovered Great Notch, and Lake Passaic lowered 80 feet (24 m) from the Moggy Hollow to the Great Notch stage. This allowed the Butler outflow to erode the north end of the Moggy Hollow delta and adjoining fans and redeposit the sediment in the Great Notch stage delta. These relationships fix the geometry of ice margin M1 (fig. 3) between Great Notch and the Highlands.

The thick, permeable fan deposits here are a prolific aquifer. Production wells for Pequannock Township (circled dots on fig. 11) are screened between 120 and 200 feet (37 and 61 m) in depth and, when first drilled, yielded between 700 and 800 gpm. The fan deposits here comprise the full thickness of the valley fill and so are unconfined, but fan gravels are more commonly buried, and confined, by lake clays (Qpml on fig. 11). Such buried, confined fan gravels are the principal valley-fill aquifers in New Jersey.

91.8	0.0	Turn left onto West Parkway from Foothills Park lot.
92.4	0.6	Turn right at stop sign, then left onto turn lane for Newark-Pompton Turnpike, then right onto NJ 23 south at light (requires crossing Newark-Pompton Turnpike).
92.5	0.1	Turn right at light onto NJ 23 south, as described above.
93.2	0.7	Flat plain here is top of a broad shallow-water delta deposited in the Totowa stage of Lake Passaic (fig. 7). Late Wisconsinan overdeepening in this area extends to more than 100 feet (30 m) below sea level; the overdeepening is filled with up to 300 feet (91 m) of recessional lacustrine deposits.
95.7	2.5	Climb from delta onto till hill.
98.0	2.3	Descend onto a stream terrace deposited in the Totowa stage, on top of lake clay.
98.5	0.5	Move left for I-80 east.
98.9	0.4	Proceed straight for I-80 east.
99.1	0.2	Exit left to I-80 east.
101.3	2.2	Roadcut through Great Notch-stage lacustrine fan which was the dam for the Totowa stage.
101.9	0.6	Cross Passaic River, much larger than before because it now includes the Pompton River.

102.4   0.5   Exit right onto Squirrelwood Road (exit 56A).

102.6   0.2   Ramp joins Squirrelwood through light.

103.1   0.5   Turn left, then left again, onto Mountain Ave.

103.4   0.3   Turn right into Garrett Mountain Reservation.

103.5   0.1   Bear right around one-way park loop road.

104.2   0.7   Bear around to left on loop road.

105.1   0.9   Turn right into parking lot for overlook. STOP 5.

Stop 5. Preglacial Drainage, Paterson Gap Overlook, Garrett Mountain Reservation

The Paterson gap in First Watchung Mountain is one of six paired wind gaps through the three Watchung ridges that, together with Sparkill gap in the Palisades ridge to the northeast (visible on the horizon from the overlook), define the preglacial route of the Hudson River in the Pliocene (fig. 1). These gaps are all between 1 and 1.5 mile (2 and 3 km) wide and have preglacial rock floors that decline evenly from 230 feet (70 m) at Sparkill to 180 feet (55 m) at Millburn, the southernmost gap. These rock-floor elevations are on grade with the Pensauken plain, tying them into a dated depositional chronology. This chronology indicates that, in the Pliocene before the pre-Illinoian glaciation, the Hudson was a tributary to the trunk Pensauken River that flowed southwesterly along the inner edge of the Coastal Plain, and discharged to the Atlantic across what is now the Delmarva Peninsula (fig. 1). The pre-Illinoian glacier diverted the Hudson and the Pensauken rivers to the Atlantic across western Long Island, and the gaps were abandoned. After the diversion, in the early and middle Pleistocene, the Passaic established a course that exited the Watchung crescent via the Short Hills and Millburn gaps (fig. 8A). The Illinoian glacier filled the Short Hills gap with till to a height sufficient to divert the Passaic northward through the Little Falls and Paterson gaps, which were not filled, in post-Illinoian time. The Passaic reestablished this route after the late Wisconsinan glaciation, which had further filled the Short Hills gap. Passaic Falls, a 77-foot drop over basalt in the center of Paterson, formed after the late Wisconsinan deglaciation as the river reestablished its course across a new escarpment between the resistant basalt floor of the wind gap and a glacially scoured siltstone lowland to the east. Hydropower from the falls was the original source of industrial development in Paterson, starting in 1791 with the "Society for the Establishment of Useful Manufactures" founded by Alexander Hamilton, which was one of the earliest industrial developments in the United States. The original design of water-power canals and city streets was by Pierre L'Enfant, who also designed the layout of Washington, D. C. Electricity is still generated at the falls.

105.1   0.0   Turn right from parking lot onto park loop road.

105.6   0.4   Turn right to park exit.

105.7   0.1   Turn left at stop sign onto Mountain Ave.

106.1   0.4   Turn left (actually, go straight) at stop sign onto Rifle Camp Road.

108.0   1.9   Turn left onto ramp for US 46 east.

108.4 0.4 Area of the Great Notch spillway, now defaced by road and railroad cuts.

108.7 0.3 Bear right to NJ 3 east.

110.1 1.4 Cross Garden State Parkway.

111.1 1.0 Cross sluiceway eroded by drainage of Great Notch stage around Garrett Mountain.

112.7 1.6 Descend onto flat surface of delta deposited in Lake Delawanna, a local lake held in by a sediment dam in the lower Passaic valley.

113.1 0.4 Cross NJ 21.

113.6 0.5 Cross Passaic River, tidally influenced here.

114.4 0.8 Roadcuts here are in a till on sandstone upland.

115.0 0.6 Cross NJ 17. Descend to the Meadowlands salt marsh, on the floor of Lake Hackensack. A narrow, glacially overdeepened trough extending to more than 300 feet (90 m) below sea level runs along the west edge of the Meadowlands and is filled with lacustrine fan deposits overlain by lake clay.

115.6 0.6 Cross Berrys Creek, move right for exit to NJ Turnpike. Giants Stadium on left, home to the NY [sic] Giants and NY [sic] Jets, and site of the 2014 Super Bowl.

116.1 0.5 Exit right to NJ Turnpike.

116.8 0.7 Bear right onto NJ Turnpike south after tolls.

118.4 1.6 View of Snake Hill, a diabase intrusion into siltstone, to left, and landfill mountains to right, resting on salt marsh atop Lake Hackensack clays.

121.6 3.2 Cross Passaic River. Newark to right, Pulaski Skyway to left.

122.2 0.6 Cross under Pulaski Skyway, completed in 1932. Depth to shale bedrock ranges from 30 to 130 feet along the 4-mile length of the viaduct. Lake clay overlies the bedrock.

129.5 7.3 Exit right at exit 13 for I-278 east and Goethals Bridge.

130.0 0.5 Bear right after NJ Turnpike tolls to I-278 east and Goethals Bridge.

132.2 2.2 Bear left onto I-278 east after Goethals Bridge tolls.

133.2 1.0 Move right for exit 8 to Victory Blvd.

133.8 0.6 Exit right to Victory Blvd. (exit 8).

134.1 0.3 Turn left at light at end of ramp onto Victory Blvd. and then move right.

134.2 0.1 Turn right into entrance to College of Staten Island.

END OF TRIP.