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Glacial aquifers of the New Jersey Highlands

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ABSTRACT

Glacial aquifers in the New Jersey Highlands are valley-fill deposits of sand and gravel laid down in glacial lakes and glacial river plains during the Illinoian and late Wisconsinan glaciations. In places they are interbedded with less-permeable silt, clay, fine sand, and till, which act as confining or semi-confining layers. The distribution of sediments within a valley-fill aquifer is determined by the geologic setting of the valley and on the volume of glacial sediment it received. There are four types of valley fill. Three (filled lake basins, unfilled lake basins, and stacked valley fills) are in valleys that contained glacial lakes and one (fluvial valley fills) is in valleys that did not contain glacial lakes. Filled lake basins have an upper unconfined sand and gravel overlying a confining or semiconfining layer of silt, fine sand, and clay, in turn overlying a lower, confined sand and gravel aquifer. Unfilled lake basins are similar but lack the upper unconfined sand and gravel. Stacked valley fills contain multiple layers of sand and gravel, silt and clay, and till. Fluvial valley fills have just a single unconfined sand and gravel aquifer. Where glacial sediments are too thin to be productive aquifers themselves, they nevertheless provide storage to feed stream baseflow and recharge underlying bedrock aquifers. Large areas on uplands north of the late Wisconsinan terminal moraine are exposed bedrock without a mantle of glacial sediment and so have little groundwater storage potential.

INTRODUCTION

The Highlands are a critically important surface–water source for northern New Jersey. Recognition of the importance of this resource has spurred conservation efforts since the Newark and Jersey City reservoir systems were built in the late nineteenth century, and is the principal scientific basis for the Highlands Water Protection and Planning Act of 2004. Maintaining the quality and quantity of streamflow is therefore a paramount management goal. Streamflow includes both runoff and baseflow. Runoff is determined in part by the infiltration capacity of surficial geologic material, and baseflow is supplied from groundwater reservoirs. In both cases, knowing the distribution and hydrologic properties of glacial sediment is essential for a full understanding of the surface-water resource. Additionally, in valleys filled with a sufficient thickness of glacial sand and gravel, the glacial deposits themselves are productive groundwater reservoirs that provide water for a number of Highlands communities.

This paper will review the history of glaciation in the New Jersey Highlands, describe the deposits that the glaciers laid down, examine the manner in which these sediments are arranged in valley-fill aquifers, and discuss how they affect runoff and streamflow. The observations made here are based on geologic mapping conducted since 1983 by the New Jersey Geological Survey. The mapping includes compilation of records of wells and borings that provide control for bedrock-surface topography and valley-fill stratigraphy. Results of the mapping are published at 1:100,000 scale for the entire Highlands (Stanford and others, 1990; Stone and others, 2002) and at 1:24,000 quadrangle scale for most of the Highlands north of the terminal moraine (maps listed at www.nj.gov/dep/njgs/pricelst/geolmapquad.htm). The quadrangle maps provide logs of wells and borings, contours of rock-surface elevation, and cross sections of valley fills. Hydrogeologic studies of valley-fill aquifers have been completed for the Lamington valley (Nicholson and others, 1996), Rockaway valley (Canace and others, 1993; Schaefer and others, 1993; Dysart and Rheaume, 1999), and the Ramapo valley (Vecchioli and Miller, 1973; Hill and others, 1992).

GLACIAL HISTORY

The New Jersey Highlands are at the southernmost limit of the area covered by the Laurentide ice sheet. The Laurentide ice sheet expands outward from accumulation centers in northern Quebec and Labrador. It has grown and melted away about 10 times within the past 2 million years. The Highlands were glaciated by the Laurentide at least three times within this period. It is possible that additional advances entered the Highlands, but their deposits have been eroded by subsequent glaciations and no evidence of their presence survives. This section will describe the extent, age, deposits, and landscape effects of these three glaciations.

Pre-Illinoian Glaciation

The oldest glaciation is known as the pre-Illinoian. It covered the entire Highlands, except, possibly, for a small part of Musconetcong Mountain in the extreme southwest (fig. 1). Deposits of pre-Illinoian age are deeply eroded and intensely weathered. They are preserved only as remnant patches, generally less than 20 feet thick, on flat hilltops and divides where they have been protected from erosion. The most extensive deposits are on flat to gently sloping terrain on carbonate bedrock in the Pohatcong, Musconetcong, and Long valleys. In these valleys the pre-Illinoian deposits lie on remnants of the former broad valley floor. Modern stream channels are inset in inner valleys that are as much as 200 feet below the pre-Illinoian valley floor. The old glacial deposits are preserved in this setting because carbonate bedrock is highly permeable due to solution channeling, and most surface water on carbonate rock drains via underground channels rather than by surface runoff. Thus, under natural conditions, there is little erosion of surface material by gullying or slopewash. This is not the case on gneiss, quartzite-conglomerate, and shale, which are the other common rock types in the Highlands, and so there is much less preservation of pre-Illinoian deposits outside the carbonate valleys. On gneiss uplands, there are some patches of pre-Illinoian till on the flat top of Schooleys Mountain and a few very small patches on flat cols and saddles elsewhere.

Most of the pre-Illinoian deposits are till, an unsorted, nonstratified deposit laid down directly by glacial ice. In the Pohatcong valley and on the southern edge of the Highlands there are a few small deposits of pre-Illinoian sand and gravel that were laid down in glacial river plains and glacial lakes. In all the deposits, pebbles and cobbles of sandstone and gneiss are fully decomposed or have thick (>0.5 inches) weathering rinds because feldspar minerals have altered to clay. Clasts of carbonate rock are fully decomposed to yellow silt due to dissolution of the carbonate minerals. Quartzite and chert clasts are generally intact and hard because they are composed almost exclusively of quartz, which does not weather easily. Some are easily broken with a hammer owing to incipient weathering along fractures and grain boundaries and most have a surface stain of iron oxide. Matrix material in the till is typically a yellowish red to reddish yellow silty sandy clay. The pre-Illinoian till has significantly more clay and is redder in color than younger glacial deposits. The clay is from weathering of feldspar minerals. The red color is from accumulation of iron oxides and hydroxides that are produced by weathering of mafic minerals.

The intensity of weathering in the pre-Illinoian deposits, and their patchy preservation and topographic position, indicate a late Pliocene or early Pleistocene age (Stanford, 2000a). In the time since pre-Illinoian glaciation, streams have cut valleys into bedrock as much as 200 feet deep, in contrast to the more recent glacial deposits, which rest within modern valley bottoms. There are two additional lines of evidence concerning age. The pre-Illinoian deposits in New Jersey correlate westward to magnetically reversed glacial deposits in central and eastern Pennsylvania (Gardner and others, 1994; Sasowsky, 1994), indicating an early Pleistocene (pre-788 ka) or older age. Pollen from a depth of 46-60 feet in a 60-foot core of lake sediment in Budd Lake contained 64% pre-Pleistocene taxa (Harmon, 1968). Budd Lake lies just outside the late Wisconsinan and Illinoian glacial limits, and the pre-Pleistocene pollen are contained in a finely laminated clay that may have been laid down in a lake dammed by pre-Illinoian deposits (Stanford and Witte, 1997). If this is so, then the pre-Illinoian deposits may be of late Pliocene age. The earliest Laurentide glaciation to reach as far south as the United States is dated by volcanic ash stratigraphy in the Missouri River valley to about 2.1 Ma (the pre-Illinoian K glaciation of Richmond and Fullerton, 1986). The pre-Illinoian deposits in New Jersey may be this same glaciation.

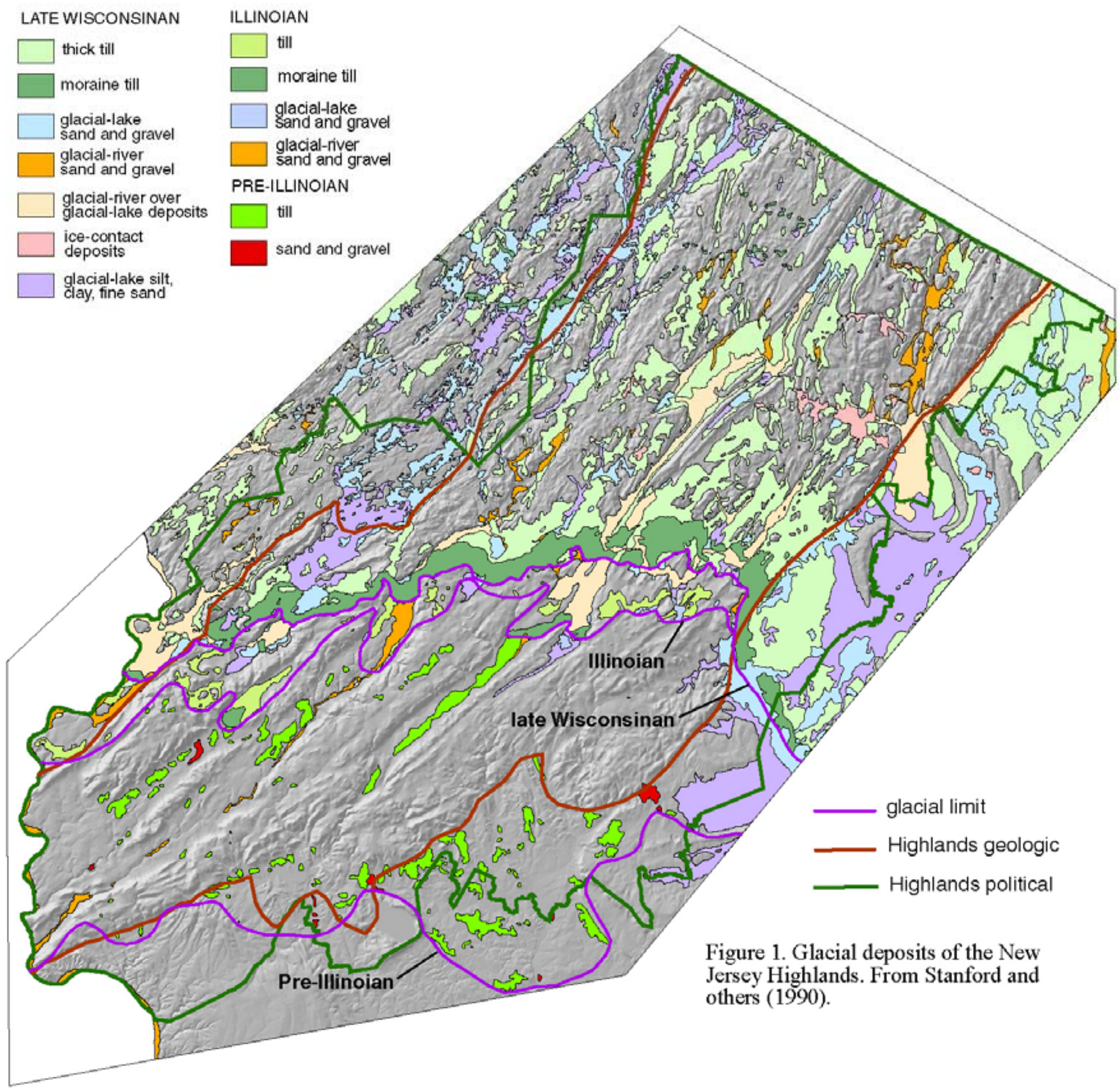

Figure 1. Glacial deposits of the New Jersey Highlands. From Stanford and others (1990).

Illinoian Glaciation

Illinoian deposits crop out in a belt about 4 miles wide south of the late Wisconsinan limit to the west of Morris Plains. They also occur in valley fills beneath late Wisconsinan deposits in the vicinity of the terminal moraine in the Lamington, Rockaway, Musconetcong, and Pequest valleys. North of the terminal moraine, Illinoian till also occurs in the cores of some drumlins, beneath late Wisconsinan till. The limit of Illinoian glaciation is marked by moraines in the Lamington valley, Long Valley, and the Pohatcong valley, and by the outer limit of Illinoian till on intervening uplands. The limit is noticeably embayed by topography, with ice lobes extending several miles down the Delaware, Pohatcong, Musconetcong, and Long-Lamington valleys. This geometry suggests a lower ice-surface profile (a rise of about 200-300 feet per mile) than that of the late Wisconsinan glacier (about 400 feet per mile). The Illinoian terminal moraine is also much smaller than that of the late Wisconsinan glaciation, a difference that is too large to be attributed to greater erosion. This difference indicates that Illinoian ice did not remain at its terminal position as long as did late Wisconsinan ice. The brief occupation time might also explain the embayed geometry of the Illinoian limit, as ice did not have time to thicken and overtop ridges.

Unlike the pre-Illinoian deposits, Illinoian deposits fill valley bottoms and are preserved on gently sloping upland surfaces. Illinioan glacial-lake deposits are as much as 250 feet thick in the valley fills north of Budd Lake and in the Lamington valley, and as much as 100 feet thick in the Musconetcong and Rockaway valleys. Original depositional landforms are preserved in places. For example, paired moraine ridges are preserved in subdued form near Washington in the Pohatcong valley, near Flanders in Long Valley, near Ironia in the Lamington valley, and north of Shongum Lake in the Rockaway valley (places named are shown on fig. 2). The persistence of double ridges in widely separated locations indicates that the ridges are an original depositional feature marking successive ice margins rather than erosional remnants of once-larger moraines. There is also a well-preserved glacial-river plain at Flanders, a well-preserved glacial-lake delta at Ironia, and somewhat more eroded but still recognizable deltas in several valleys north of Shongum Lake.

Weathering is less intense than in pre-Illinoian deposits. Matrix color is brown and yellowish brown rather than reddish yellow. Gneiss and sandstone pebbles and cobbles have thin weathering rinds (<0.25 inches) but are otherwise intact. Both these features indicate only incipient weathering of feldspar and matrix minerals. Carbonate clasts, however, are entirely decomposed to yellow silt, as in pre-Illinoian deposits, owing to the rapidity of carbonate-mineral dissolution.

The distribution of meltwater deposits indicates that glacial-river plains were laid down in the Delaware, Musconetcong, and Drakes Brook-South Branch Raritan River valleys during Illinoian glaciation. All of these valleys drain southward, away from the glacier, and so meltwater could drain freely down them. The upper Lamington valley, which at the time was a tributary to the Rockaway, and the Mill Brook and Den Brook valleys in the Rockaway basin around Lake Shongum, drained northward and so were dammed by the ice margin. Lake Ironia formed in the upper Lamington valley, and Lake Shongum formed in the Mill and Den Brook valleys (fig. 2). Lake Ironia spilled southward down the lower Lamington into the Raritan basin, and Lake Shongum spilled southward into the Whippany valley. Illinoian glacial lakes also occupied parts of the Rockaway, Musconetcong, and Pequest valleys, and a former north-draining valley north of Budd Lake, as indicated by thick glacial-lake deposits of interbedded sand and gravel and silt, clay, and fine sand that are buried beneath late Wisconsinan deposits in those valleys. The elevation, extent, and drainage history of these lakes cannot be fully determined owing to erosion and burial of the deposits, but they likely were similar to the late Wisconsinan lakes Oxford, Budd, Waterloo, Dover, and Denville.

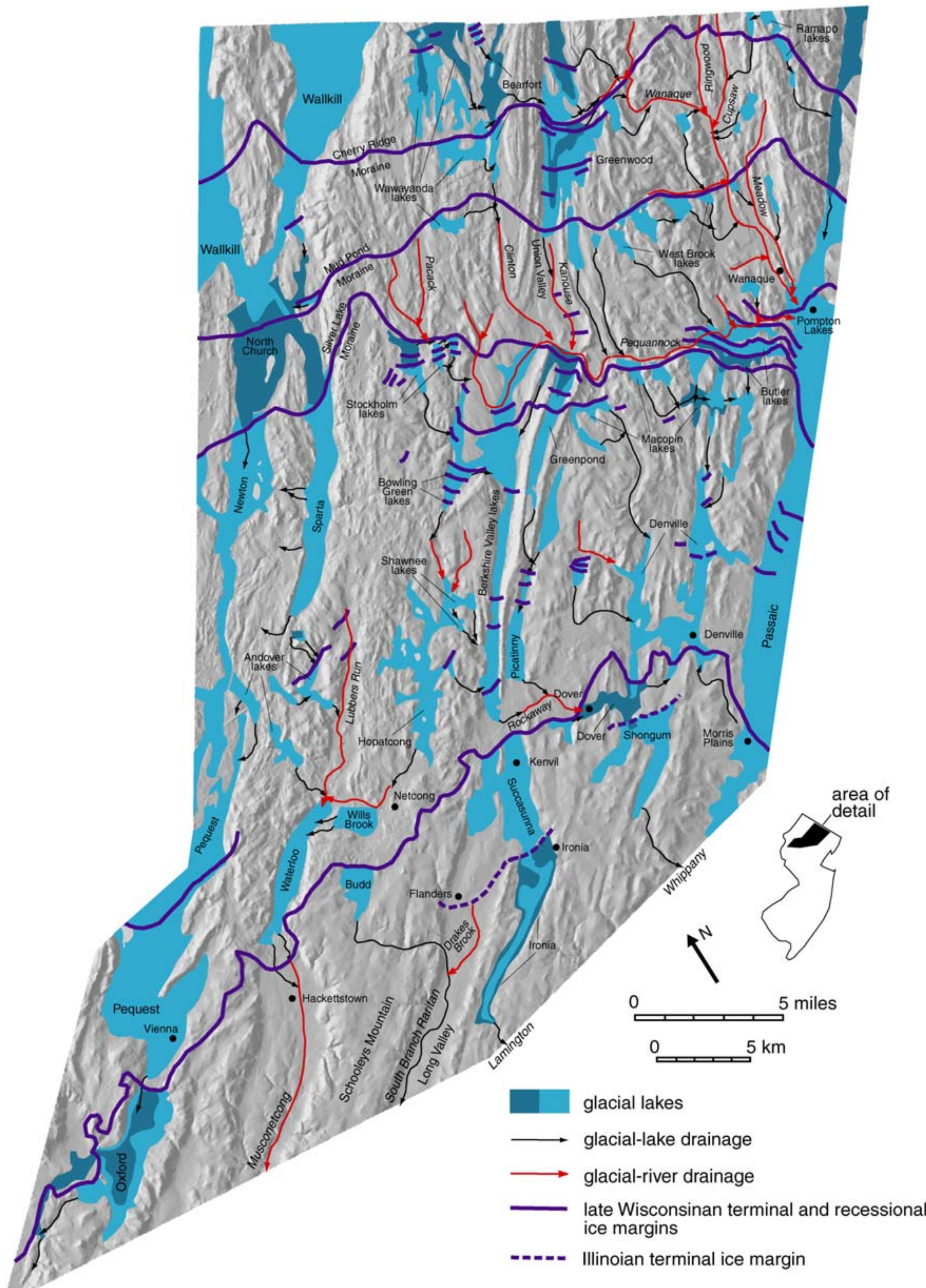

Figure 2. Late Wisconsinan and Illinoian glacial lakes, ice margins, and meltwater drainage.

The Illinoian deposits have not been directly dated. East of New Jersey they correlate to till on Nantucket that is overlain by marine deposits that are about 125,000 years old (oxygen-isotope stage 5) (Oldale and others, 1982). The Illinoian glaciation (oxygen-isotope stage 6, about 150 ka) immediately preceded the stage 5 marine highstand, and has an extent similar to that of the late Wisconsinan glacier, based on the amplitude of the marine oxygen-isotope record at this time. Thus, it is the most likely age of the till. Soil development and weathering intensity in Illinoian deposits in New Jersey are similar to that of the type Illinoian tills in the midwestern United States. Some workers have considered these deposits, and correlative deposits in Pennsylvania, to be early Wisconsinan (about 70 ka) in age (Sevon and others, 1975). Early Wisconsinan glacial deposits occur in the Great Lakes basin but sea-level and marine oxygen-isotope records indicate that there was insufficient ice volume at this time to overtop the Allegheny Plateau and bring ice south to New Jersey. Also, the degree of weathering and erosion is excessive for an early Wisconsinan age (Ridge and others, 1990).

Late Wisconsinan Glaciation

Late Wisconsinan deposits and erosional landforms cover the entire Highlands north of the late Wisconsinan limit (fig. 2). In valleys along the terminal moraine, and in the cores of a few drumlins, ice did not erode deeply and Illinoian deposits and weathered bedrock are preserved beneath late Wisconsinan deposits. Nearly everywhere else, the ice eroded all previously existing surficial material and the late Wisconsinan deposits lie directly on bedrock. In a few spots, preglacial gneiss saprolite and carbonate-rock residuum as much as 200 feet thick are preserved beneath till. These are remnants of thick or narrow, deep zones of weathered rock that were not entirely removed by glacial scour.

Unlike the older deposits, late Wisconsinan sediments show little weathering. Gravel clasts of all lithologies except carbonate rock are unweathered, although till near the glacial limit includes a few percent weathered gneiss and sandstone pebbles that have been reworked from Illinoian deposits. Where carbonate-rock clasts are not abundant (<10% of total clasts), they are commonly decomposed to depths greater than 10 feet thick, but where abundant the carbonate is much less weathered. Matrix color commonly reflects that of the constituent rock fragments rather than weathering effects. Similarly, glacial landforms are only slightly modified by postglacial erosion. Significant erosion is limited to some steep slopes where the till has moved downslope to accumulate as aprons of colluvium, and to some bank erosion along larger streams. Small-scale details like ridges, basins, and knolls in moraines, and esker ridges, deltas, and fans in glacial lakes are preserved largely intact.

Radiocarbon dates on peat, organic sediment, shells, wood, and organic concretions recovered from above and below the late Wisconsinan till in New Jersey and on Long Island (summarized in Stanford, 2000) indicate that late Wisconsinan ice advanced into the New Jersey Highlands about 22 ka (in radiocarbon years before present) and had retreated north of the Highlands by about 18 ka. The orientation of striations and drumlins, and the provenance of clasts in till, indicate that ice advanced southward across the Highlands from the Kittatinny-Wallkill valley (Stanford, 1993). During advance, till was laid down in ramps and tongues on northwest-facing hillslopes that faced the advancing ice, and in drumlins. Till in ramps is as much as 100 feet thick; in drumlins, as much as 200 feet thick.

Most till in the Highlands is very pale brown to yellowish brown silty sand till with many cobbles and boulders (the Netcong Till of Stone and others, 2002) derived from erosion of the local gneiss bedrock. Pebbles and cobbles are chiefly gneiss, gray sandstone, and, in the eastern Highlands, purple conglomerate. Boulders are chiefly gneiss and conglomerate. Along the west edge of the

Highlands, yellowish brown silty till (Kittatinny Mountain Till of Stone and others, 2002), derived from carbonate rock and shale of the Wallkill and Kittatinny valleys, locally extends as tongues up to several miles into the Highlands. These till tongues, and scattered erratic boulders of carbonate rock through the western Highlands, reflect southerly ice flow. Locally in Berkshire and Union valleys there are scattered deposits of reddish silty till derived from red shale of the Longwood Shale. Much upland terrain on resistant gneiss and quartzite-conglomerate is glacially eroded bedrock outcrop with little or no till cover (refer to fig. 6 below). Northwest and north-facing slopes were abraded to form smooth, sloping ledges, while southeast and south-facing slopes were plucked to form cliffs and blocky outcrops.

Ice advanced to the east-west trending linear valley across the Highlands that is formed by segments of the Rockaway valley (between Denville and Kenvil), the Musconetcong valley (between Netcong and Hackettstown), and the Pequest valley (west of Vienna). This topographic trough acted as a pinning line and halted further advance. The ice front remained in this position for several hundred years, possibly a thousand years, and deposited the terminal moraine. The terminal moraine is a belt of till averaging about two miles wide and up to 150 feet thick, with ridge, knoll, and basin surface topography. It was built gradually by lodging of till at the base of the glacier and by sliding and melt-out of debris from the glacier surface and from detached ice blocks.

As the terminal moraine was forming, meltwater laid down stratified sediments in glacial lakes in the Rockaway valley (lakes Denville and Dover), the Lamington valley (Lake Succasunna), the Musconetcong valley (lakes Budd and Waterloo), and the Pequest valley (Lake Oxford) (fig. 2). All of these valleys drained toward the glacier margin, and so were dammed by the ice. The Musconetcong below Hackettstown, and the Delaware, slope away from the glacier and so meltwater drained freely down the valley and deposited glaciofluvial plains.

As the glacier front retreated from the terminal moraine, meltwater deposited stratified sediment in a succession of glacial lakes and river plains (figs. 2 and 3). Glacial lakes formed in valleys that, as was the case with glacial lakes at the terminal moraine, sloped toward the retreating ice margin and so were dammed by the glacier, or that were blocked by previously deposited glacial sediment. Most of the recessional lakes are of the first type. These include: Lake Newton, which was in a north-draining segment of the Paulins Kill valley; lakes Sparta, North Church, and Wallkill, which were in the north-draining Wallkill valley; Lake Bearfort, which was in the north-draining Longhouse Creek valley (a tributary to the Wallkill); the Wawayanda lakes, which are in small north-draining valleys on Wawayanda Mountain; Lake Greenwood, which is in the north draining Belcher Creek valley (a tributary to the Wanaque); the West Brook lakes, which are in north-draining tributary valleys in the Wanaque basin; the Ramapo lakes, which are in a north-draining valley on Ramapo Mountain; Lake Greenpond and the Stockholm, Macopin, and Butler lakes, all of which are in north-draining valleys on the south side of the Pequannock basin; and the Shawnee, Wills Brook, Bowling Green, and Andover lakes, which are in north-draining valleys in the Musconetcong, Musconetcong again, Rockaway, and Pequest basins, respectively.

Sediment-dammed lakes include Lake Pequest, Lake Hopatcong, the Berkshire Valley lakes, and Lake Picatinny, all dammed by the terminal moraine. The northern reaches of Lake Pequest and the Berkshire Valley lakes are dammed by delta deposits laid down in the earlier, more southerly parts of the lakes dammed by the moraine. Because these deltas include a topmost river-plain deposit that lies above lake level, the more northerly lakes have water levels that are slightly higher than the original, moraine-dammed lakes. In other cases, erosion of the sediment dams caused lake levels to lower as the ice front retreated, permitting river plains to be deposited atop drained lake beds or in shallow relict ponds. Figure 3 summarizes the temporal sequence and drainage history of these lakes.

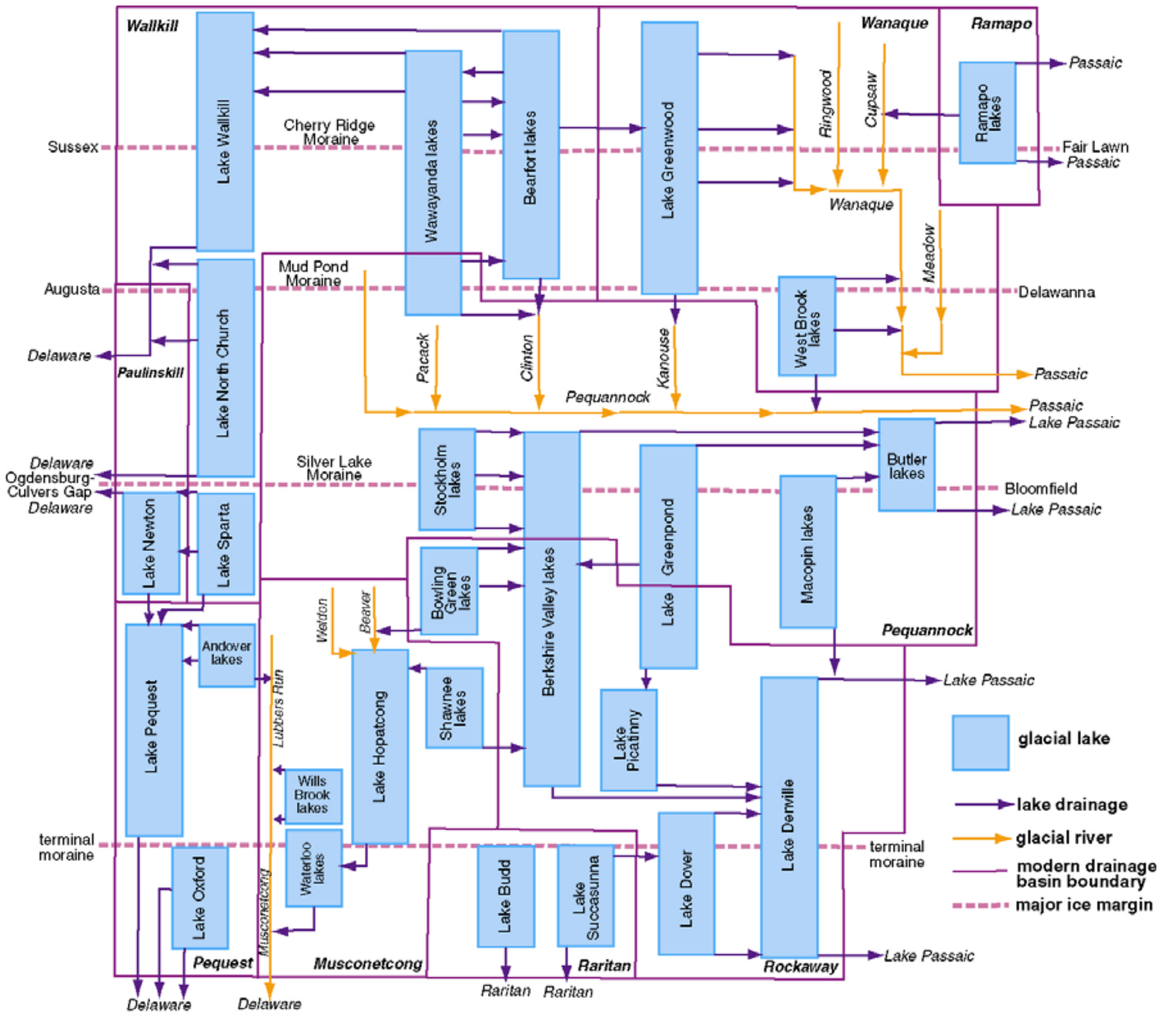

Figure 3. Correlation of late Wisconsinan glacial lakes, meltwater drainage routes, and ice margins. Modified from Stanford (1993).

The meltwater deposits are well-sorted and stratified. Gravel and sand composition closely mimics local bedrock. Glaciofluvial deposits range from cobble-to-boulder gravel close to ice-margin positions, or where glacial rivers eroded bouldery till, to pebbly sand in downstream areas. Coarse gravels are typically in thick, structureless beds. Pebbly and sandy deposits are typically channeled and cross-bedded. Glaciolacustrine sediments include sand and gravel deposited in deltas and fans, and silt, clay, and fine sand deposited on lake bottoms. Deltas are flat-topped ridges or terraces of sand and gravel built out into lakes from ice margins. They are fed by meltwater issuing from tunnels beneath or within the ice, or by meltwater draining from uplands along the edge of the ice. They typically consist of a thick section (as much as 150 feet) of inclined beds of sand and pebbly sand, known as foreset beds, that are laid down as sediment lobes prograde into the lake. The foreset beds are overlain by thin (generally less than 15 feet) horizontal beds of pebble-to-cobble gravel, known as topset beds, that are laid down as a plain atop the delta by streams draining across the delta. The contact of the topset and foreset beds marks the former lake level. Generally, deltas coarsen upward from sand through pebbly sand to pebble and cobble gravel, as a result of progradation.

Lacustrine fans are knolls and ridges of sand and gravel deposited on lake bottoms along ice margins. They are fed by meltwater issuing from tunnel channels beneath or within the glacier. They consist of interbedded pebble-to-cobble gravel, pebbly sand, and sand. Beds are typically subhorizontal or gently inclined and bounded by scour surfaces. Sands typically show planar lamination, cross bedding, and some climbing ripples, indicating rapid current velocities and high sedimentation rates. Gravels are thick bedded and massive and have little fine material, again indicating high current velocities. Meltwater in the tunnel channels flows under high hydraulic head, and velocities drop sharply when the water exits the tunnel into the lake, accounting for the rapid changes in grain size and the scour and bedding structures. As deposits aggrade at the tunnel mouth, the tunnel will shift laterally or vertically, creating a deposit with multiple ridges and knolls. Laucstrine fans are as much as 100 feet thick. Along stable ice margins, lacustrine fans may aggrade to the lake surface and then prograde into the lake as deltas.

Clay, silt, and very fine sand remain in suspension and are carried out onto the lake bottom by high-density turbidity flows. These flows travel down the delta and fan fronts and collect in the lowest spots in the lake basin. The sediment settles out and aggrades to form lake-bottom deposits. These deposits are typically horizontally laminated. Silt and fine sand laminations reflect individual turbidity flows carrying in pulses of sediment. Clay layers record winter sedimentation, when lakes are frozen over, meltwater production is minimal, and the finest clay particles settle out. These annual laminated deposits are known as varves. Lake-bottom deposits are as much as 150 feet thick.

Most sediments laid down during retreat are stratified meltwater deposits in valleys, but there are a few noteworthy till and till-like recessional deposits. Three recessional moraines (the Silver Lake, Mud Pond, and Cherry Ridge moraines, from south to north) traverse Hamburg and Wawayanda mountains (Stanford, 1993). They tie to the Ogdensburg-Culvers Gap, Augusta, and Sussex ice margins, respectively, in the Kittatinny Valley (Witte, 1997), and link eastward to the Bloomfield, Delawanna, and Fair Lawn ice-margin positions in the Newark Basin (Stanford and Harper, 1991). Like the terminal moraine, these recessional positions are pinned by deep east-west trending valleys in the Highlands east of the moraines, specifically the Pequannock valley, the West Brook valley, and the Hewitt Brook valley for the Silver Lake, Mud Pond, and Cherry Ridge moraines, respectively. Thicker ice in these valleys slowed the retreat of the ice margin. The recessional moraines are discontinuous and much smaller than the terminal moraine, but in places may have 50 feet of relief.

In several upland valleys, and in the deep Pequannock valley downstream from Kanouse Brook, there are deposits of sandy, bouldery, noncompact till-like sediment with subdued ridge-and-basin topography ("ice-contact deposits" on fig. 1). These deposits are in settings where ice masses could have become separated from the main glacier by downwasting during retreat. Sediment released from, or deposited around, the ice masses, would be noncompact and poorly sorted, and would have a hummocky landform due to slumping and collapse when the ice melted. In the Pequannock valley the deposits include collapsed delta and lacustrine-fan sand and gravel that were laid down around stagnant ice in lakes in north-draining tributary valleys on the south side of the main valley (Macopin and Butler lakes, figs. 2 and 3).

Postglacial Deposits

As ice retreated, glacial lakes drained and glacial river plains became abandoned. Postglacial streams established new courses through the glacial landscape, incising channels and floodplains into the valley-fill deposits. Some former lake bottoms in valleys, and numerous glacially scoured rock basins on uplands, remained as postglacial lakes. Many of these were gradually filled with silt, clay, and peat to form marshes and swamps; others were deeper or received little sediment and so remain as lakes. Pollen preserved in these postglacial organic sediments provide an important record of postglacial vegetation and climate. Talus accumulated, and continues to accumulate, at the base of glacially plucked cliffs. The postglacial alluvial, wetland, and talus deposits are generally less than 15 feet thick, although some of the wetland deposits are as much as 40 feet thick.

VALLEY-FILL AQUIFERS

Glacial aquifers occur where valley fills exceed 50 feet in thickness (fig. 4). Within this zone, aquifers occur where permeable sand and gravel is of sufficient saturated thickness or, in the case of confined aquifers, has sufficient pressure head, to supply usable quantities of water to wells. These conditions generally are met only in select areas of a given valley fill. However, recharge of these select aquifers, and groundwater flow paths within them, are governed by the stratigraphy of the entire valley fill, the geometry of the valley-fill sediments, and the geometry of the bedrock surface containing the valley fill.

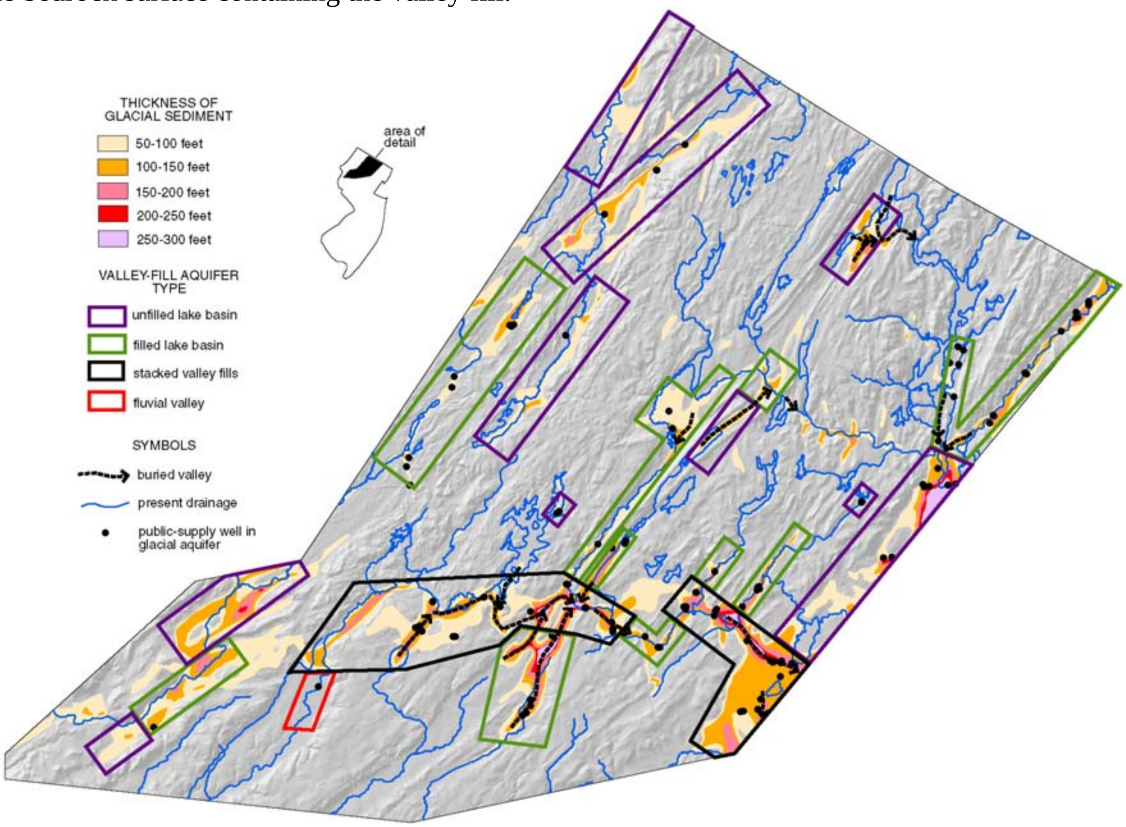

Figure 4. Valley-fill aquifers and thickness of glacial sediment.

Valley-Fill Aquifer Types

There are four basic types of valley fill in the Highlands (fig. 5). Fluvial valley fills (fig. 5a) are in valleys that drained away from the glacier, and were not blocked by sediment dams, and so did not contain glacial lakes. In the Highlands, only the Delaware, Musconetcong below Hackettstown, the Lubbers Run valley, some northern tributaries of the Pequannock, and parts of the Wanaque valley are in this category. In these valleys, plains of sand and gravel are laid down from successive ice margins. Each plain rises in altitude and coarsens in texture toward the ice margin from which it was deposited. As the ice margin retreats, the plains are eroded and reworked by meltwater draining from the up-valley ice margins. In the narrow valleys of the Highlands, this erosion tends to reduce the plains to terrace fragments along the valley sides and the reworking tends to homogenize the texture of the deposits. Because there is no lake to act as a settling basin, silt and clay remain in suspension and are flushed from the valley. The valley fill deposits are pebbly sand, pebble-to-cobble gravel, sand, and, locally, boulder gravel. Because no basin is being filled, the deposits are generally less than 50 feet thick. In places, glacial erosion during advance may have scoured rock basins in the valley floor (for example, beneath Lake Lackawanna in the Lubbers Run valley) that are then filled with lacustrine sand and silt before being covered by glaciofluvial sand and gravel. Because there is little or no fine-grained sediment, fluvial valley fills are unconfined, and there is generally excellent hydraulic connection between lakes, streams, and the valley fill. The thin fill depths, and fragmentary distribution due to erosion, limit their use as aquifers. However, shallow wells positioned to induce infiltration from streams or lakes may be highly productive. Only the fluvial valley fill on the Musconetcong near Hackettstown is sufficiently thick to warrant consideration as a potential aquifer.

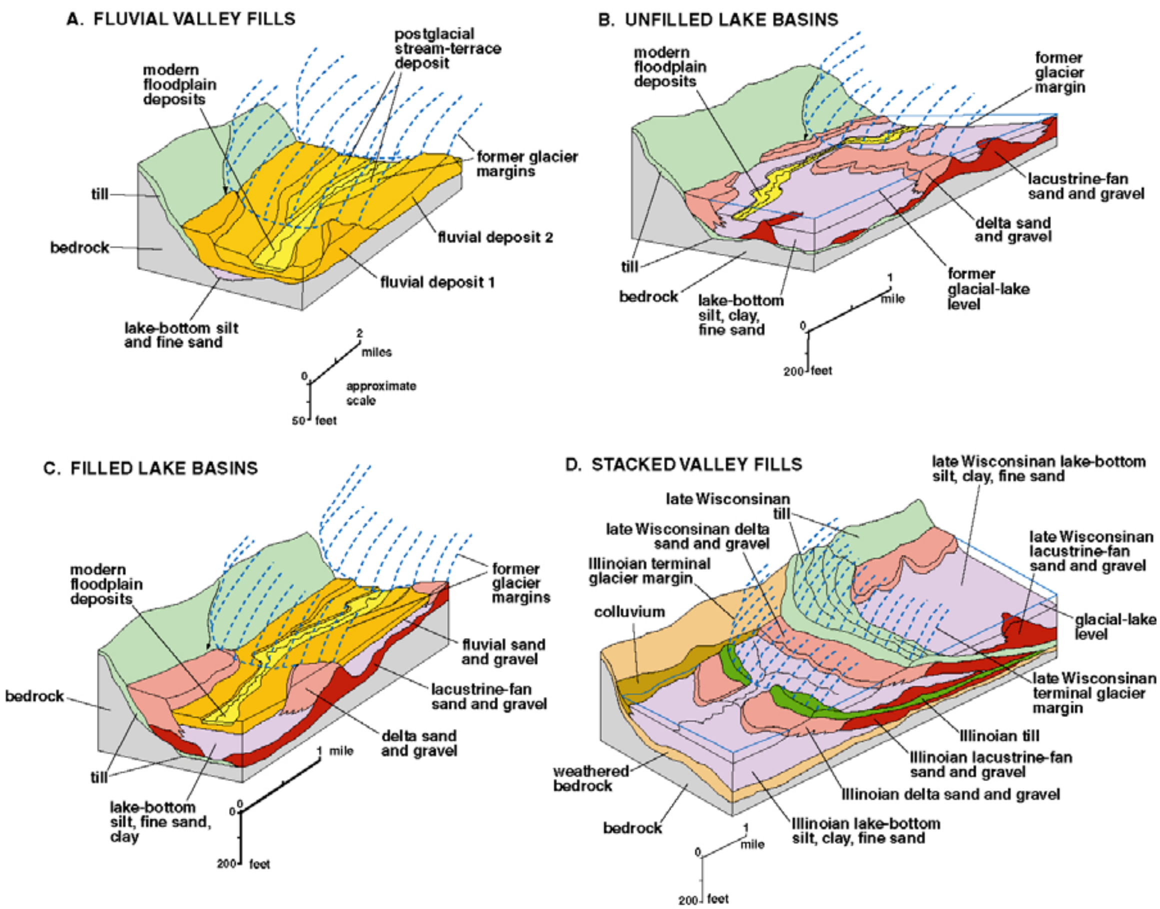

Figure 5. Types of valley-fill aquifer.

Unfilled lake basins (fig. 5b) are in valleys containing glacial lakes that were too large, or received too little glacial sediment, to fill completely with deposits. Landforms in these basins are typified by deltas and fans separated by lake-bottom plains. The lake-bottom plains are commonly occupied today by floodplains, marshes, and swamps. In this type of valley fill, lacustrine-fan deposits may be highly productive confined aquifers where they are covered by lake-bottom silt and clay. The lacustrine-fan deposits, however, are discontinuous and may be hard to locate where they are buried by lake clay. In places (for example, in the Wallkill valley) the fans were laid down sequentially at the mouths of long-lived tunnel channels, and so form linear tracks up the valley. Elsewhere, fan deposition is more random owing to ephemeral or shifting tunnel channels. Where not covered by lake clay, deltas and outcropping fans may be unconfined aquifers if sufficiently thick and saturated and may act as recharge conduits for the confined lacustrine fans. Recharge is particularly feasible where deltas with large surface extent are in contact with outcropping fans that, in turn, are continuous with confined fans. Lakes Oxford, Pequest, Sparta, North Church, Wallkill, and Passaic are all large unfilled lakes; smaller unfilled lakes include Hopatcong, Greenpond, and Greenwood.

Filled lake basins occur in valleys where lakes were small enough or received enough glacial deposition to fill completely. As in unfilled lakes, lacustrine-fan deposits occupy the valley floor, are overlain by lake-bottom deposits, and extend upward in places to connect with deltas. However, because these lakes filled or drained while ice was still releasing meltwater and sediment into the valley, there is an upper sand and gravel above the lake-bottom deposit. This upper sand and gravel was laid down as a river plain or a shallow-water delta from up-valley ice margins. Thus, there is an upper, unconfined aquifer in addition to the confined lacustrine-fan aquifer. In many cases this upper aquifer is too thin to be a significant source of water, but where it is contiguous with deltas, or with outcropping fans, there may be sufficient saturated thickness for production. As with the fluvial valley fills, there is excellent hydraulic connection between the upper aquifer and surface water. Thus, the upper aquifer may act as a recharge conduit for surface water moving into the confined aquifer. Filled lake basins most commonly occur in narrow valleys that contained small glacial lakes, such as Berkshire Valley, the upper Pequest valley, lakes Picatinny, Dover, and Denville, and the Ramapo and lower Wanaque valley. In larger valleys they occur only adjacent to major ice margins, such as Lake Succasunna and parts of lakes Oxford and Pequest, all near the terminal moraine, and the Germany Flats section of Lake Newton, which fronts the Ogdensburg-Culvers Gap moraine. In these settings, deposition was prolonged due to the long residence time of the ice margin at that position.

Stacked valley fills (fig. 5d) occur in glacial lakes along the terminal moraine. Here, late Wisconsinan ice did not erode deeply, in part because the glacier thins at its edge and in part because lake water buoyed the glacier off its bed. Preexisting Illinoian deposits were not completely eroded as they were further north, and late Wisconsinan till and lake deposits were laid down on top of similar Illinoian deposits. The valley fill thus consists of stacked layers of till, lacustrine-fan sand and gravel, lake-bottom silt, clay, and fine sand, and delta sand and gravel. Although not depicted in figure 5d, there may be a topmost layer of glaciofluvial sand and gravel, laid down after lakes drained or filled and ice retreated up-valley. The lake-bottom deposits and, to a lesser degree, till, are confining beds. Till in the Highlands has a silty sand matrix but the matrix is generally compact, reducing the permeability. As in filled lake basins, the lacustrine-fan beds may be confined aquifers and the topmost delta and river-plain deposits are unconfined aquifers. In the Highlands, stacked valley fills occur along the terminal moraine in the Musconetcong valley between Hackettstown and Lake Hopatcong, in the buried valley beneath the terminal moraine north of Budd Lake, in the upper Lamington valley, and in the Rockaway valley. In most of these valley fills the principal aquifer is thick lacustrine-fan sand and gravel laid down during retreat of Illinoian ice. Overlying late Wisconsinan deposits include till, lake-bottom silt and fine sand, and unconfined delta and glaciofluvial sand and gravel.

Flowpaths

Groundwater flow in the valley-fill aquifers depends in detail on the distribution of permeable and impermeable sediment within the valley fill, land-surface topography, topography of the bedrock surface, and location of lakes and streams. As a general rule, flow in unconfined aquifers conforms to land-surface topography, with flowpaths trending down-valley and towards the main stream in valley bottoms. In confined aquifers, flow may follow the axis of the bedrock valley or trough containing the valley fill, which in places differs from land-surface topography and modern stream drainage owing to glacial drainage dislocations. These buried valleys are shown by arrowed, dashed lines on figure 4. For example, flow in the lower confined aquifer in the Rockaway valley fill follows a buried valley east of Denville, where the modern Rockaway River turns north (Schaefer and others, 1993). Flow in the upper unconfined aquifer, in contrast, follows the present river. Similarly, flow in the confined aquifer in the Lamington valley fill is generally northward, along the slope of the buried valley, although the potentiometric surface is altered from its natural condition in places by well pumpage (Nicholson and others, 1996). Flow in the unconfined aquifer is southward, following the modern drainage of the Lamington River. Similar divergence of confined-aquifer and unconfined-aquifer and surface-water flow likely also occurs in buried valleys north of Budd Lake and Green Pond. Both are valleys that drained north before the Illinoian glaciation but that now drain south after they were filled with glacial deposits. In both cases, confined flow crosses modern drainage-basin divides. This likely also occurs in confined sand and gravel in the buried valley southeast of Lake Hopatcong, which is a former tributary to the Rockaway that crosses the modern Rockaway-Musconetcong divide. In several other cases, such as the Rockaway at Dover, the Wanaque south of Greenwood Lake and between Wanaque and Pompton Lakes, and the Ramapo at Pompton Lakes, confined groundwater flow deviates from that of the surface drainage within a basin. In these cases, segments of the valleys were filled with glacial deposits and postglacial streams are shifted into former side valleys or across former bedrock ridges.

Vertical movement of water with a valley fill depends on the thickness, lateral extent, and permeability of fine lake-bottom sediment and, in stacked valley fills, till. Thick deposits of silt and clay and compact, silty till are effective confining layers and will impede vertical movement of water from unconfined aquifers, streams, and lakes into confined aquifers. Where such confining layers are present, water must move laterally into confined aquifers from surface sources where the confining layer is absent. For example, lake-bottom deposits are generally absent along former ice margin positions where deltas connect downward to lacustrine fans, along valley walls where sandy till or deltas connect to buried lacustrine fans on the valley bottom, or where buried lacustrine fans crop out. Where lake-bottom deposits contain abundant fine sand, as is the case in many of the filled lake basins, and where till is sandy, the confining unit is leaky and will permit some direct vertical flow from the surface into the lower aquifer.

Field and laboratory tests indicate that the hydraulic conductivity of aquifer sand and gravel in the Highlands ranges from 350 to 2600 ft/d (Gill and Vecchioli, 1965; Vecchioli and Miller, 1973; Hutchinson, 1981; Canace and others, 1983; Hill, 1985; Sirois, 1986; Hill and others, 1992; Nicholson and others, 1996; values summarized in Stanford, 2000b). Laboratory tests on three samples of fine-sandy clayey silt lake-bottom sediment from the Ramapo and Lamington valley fills yielded values of $3x10^{-4}$, $3x10^{-4}$, and $4x10^{-3}$ ft/d (Hill and others, 1992; Nicholson and others, 1996). Clayey silt to silty clay lake-bottom deposits are probably one or two orders of magnitude less conductive than fine-sandy silt. Till has not been tested, but hydraulic conductivities in the range of 10^{-3} to 10^{0} ft/d are likely for the sandy silt to silty sand tills in the Highlands. This range is confirmed by modeling results in the Dover area (Dysart and Rheaume, 1999).

Recharge

The valley-fill aquifers are recharged by: 1) direct precipitation on outcropping parts of the aquifer, 2) infiltration of water from streams and lakes into the aquifer, and 3) flow of groundwater from adjacent bedrock. Direct precipitation is significant for aquifers with large areas of sand and gravel outcrop. In the Highlands, large deltas in the southern end of Lake Pequest, in Lake Newton, Lake North Church, and Lake Succasunna provide extensive sand and gravel outcrop for collecting precipitation. Infiltration of water from streams and lakes is particularly important in narrow valleys with small areas of aquifer outcrop that are traversed by large streams. Such conditions typify most Highland valleys, particularly the Rockaway and Ramapo valleys. Streamflow measurements in these valleys demonstrate losses from the Rockaway and Ramapo rivers adjacent to pumping wells (Vecchioli and Miller, 1973; Schaefer and others, 1991; Hill and others, 1992; Dysart and Rheaume, 1999). Modeling results, streamflow measurements, and field observations in the Lamington and Ramapo valleys also demonstrate losses on tributary streams as they flow from uplands onto sand and gravel in the valley, indicating that they provide recharge to the valley fill (Hill and others, 1992; Nicholson and others, 1996). Elsewhere, or during wet periods, streams may gain flow from the valley-fill deposits. This is particularly true of the main stream in the valley, which flows at the lowest elevation and is generally incised in the valley-fill deposits.

Recharge from bedrock is more difficult to measure, but modeling results and aquifer tests demonstrate good hydraulic connection between solution-channeled carbonate rock in the Pequest and Lamington valleys (Hutchinson, 1981; Nicholson and others, 1996) and overlying sand and gravel deposits, indicating that water moves freely between the two aquifers. Gneiss, quartzite, and conglomerate are much less permeable than carbonate rock, except where weathered, and so probably do not contribute significant recharge to adjoining valley fills.

Stream Baseflow

Only a small portion of the Highlands is underlain by valley-fill deposits. The vast majority of the Highlands are uplands of gneiss or, in places, quartzite and conglomerate, bedrock. South of the terminal moraine the gneiss is mantled by weathered rock material, including clayey sand to sandy clay saprolite and blocky rock rubble, that may be as much as 100 feet thick but is more generally 20 to 30 feet thick. Lower parts of long hillslopes are covered by aprons of colluvium, which is weathered rock material that has moved downslope by creep, solifluction, and slopewash. Colluvium may be as much 50 feet thick. Outcrop is restricted to a few narrow ridgetops, steep slopes, and ravine banks. North of the moraine, glacial erosion removed almost all of the weathered-rock material and colluvium, exposing the underlying bedrock (fig. 6). Till was deposited atop the rock in places, but over much of the upland area till is patchy and less than 15 feet thick (gray areas on fig. 6), and many areas have virtually no surficial material (black on fig. 6). Locally, sandy till more than 50 feet thick, or granular weathered rock preserved beneath till, is thick and permeable enough to supply domestic wells. These thick deposits are present only in the cores of a few drumlins, and in till ramps on north- or northwest-facing slopes (light gray patches outside valleys on fig. 4).

However, even where till deposits are not thick enough to be aquifers, they serve an important hydrologic role as groundwater storage that supplies baseflow to streams. Sandy till has a porosity between about 25 and 35% (Melvin and others, 1992) whereas gneiss has a porosity of

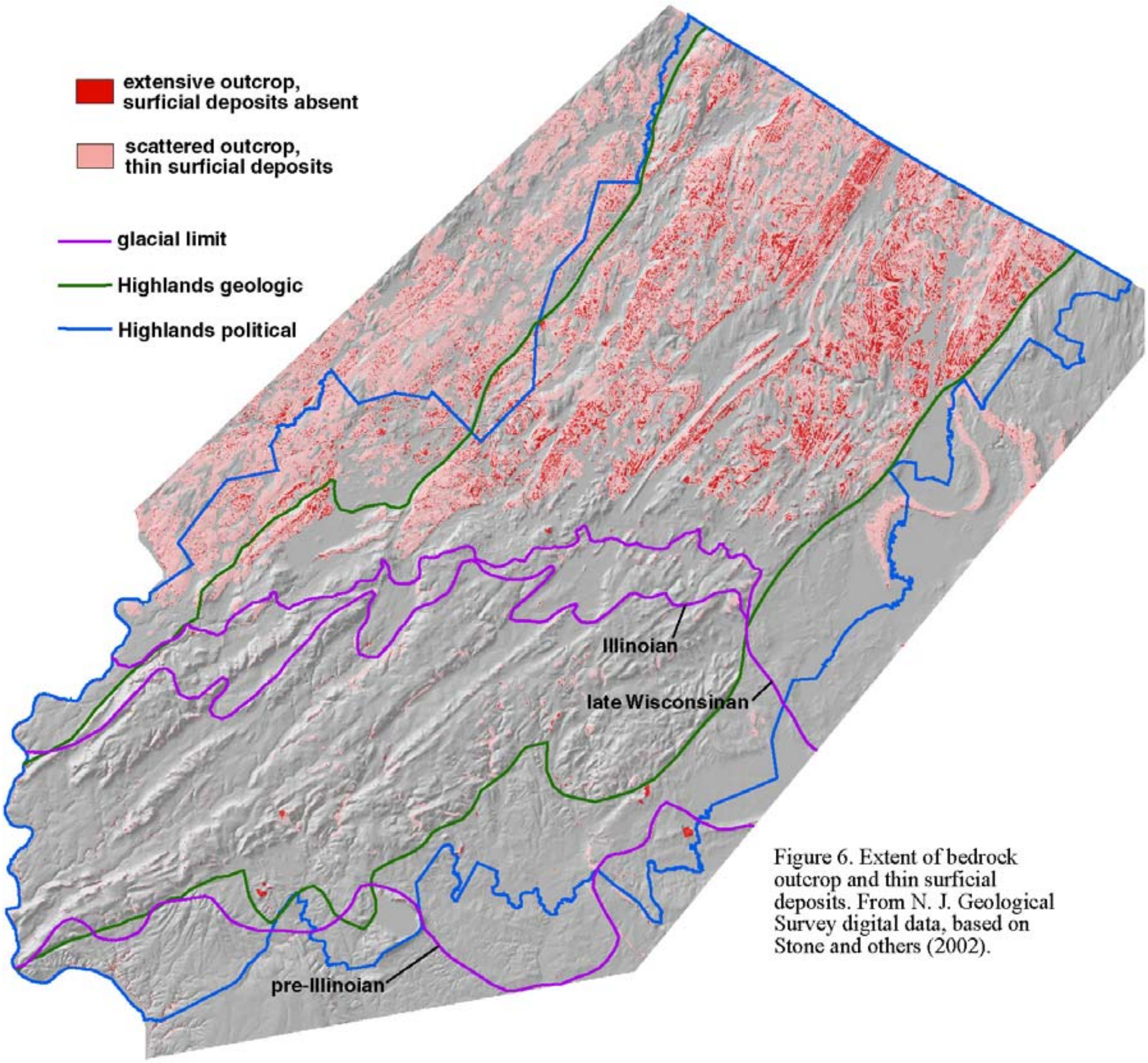

Figure 6. Extent of bedrock outcrop and thin surficial deposits. From N. J. Geological Survey digital data, based on Stone and others (2002).

about 2% (Randall and others, 1966). Till is thus about 10 to 20 times more porous than the underlying gneiss. More rainfall and snowmelt will infiltrate till than gneiss (or quartzite) bedrock. Till that mantles lower slopes and upland valley bottoms, which is the usual landscape position of till on uplands in the Highlands, will absorb runoff from upslope rock outcrops. Although no comprehensive studies of factors influencing baseflow have been conducted in the Highlands, research in similar glaciated terrain in New England can be used as a guide to conditions here. Low flows per square mile of drainage basin of streams in New England are much higher in basins containing sand and gravel than in basins containing till and bedrock (Wandle and Randall, 1994). Baseflow differences between till basins and bedrock basins have not been evaluated, but are likely similar to, or greater than, those between valley fill and till, given the similar, or greater, difference in porosity and permeability between the two materials.

Seepage of groundwater from lower parts of till-mantled slopes is commonly observed in the field, in contrast to the dry condition of rock slopes. This observation also demonstrates the groundwater storage capacity of till. Given the vital importance of surface water in the Highlands, the geologic, biologic, and land-use factors affecting runoff, seepage, baseflow, and the exchanges between groundwater, lakes, streams, and wetlands is a subject worthy of detailed study.

A CASE EXAMPLE: THE RAMAPO VALLEY-FILL AQUIFER

The Ramapo valley forms the northeast border of the New Jersey Highlands. Although half is outside the geologic boundary of the Highlands it is nevertheless typical of filled lake basins throughout the Highlands. It is a narrow valley bordered on the northwest by gneiss underlying Ramapo Mountain, and on the southeast side by Jurassic and Triassic basalt, sandstone, and shale, which is covered by as much as 150 feet of till in places. Shale and sandstone underlie most of the valley floor. The gneiss is separated from the Jurassic and Triassic rocks by the Ramapo Fault, which underlies the northwest side of the valley. Late Wisconsinan ice within the valley advanced to the southwest and scoured the shale and sandstone in the valley bottom. Topography of the bedrock surface shows that the valley bottom has been overdeepened by as much as 150 feet compared to its preglacial elevation (Stanford, 2004). The valley-fill deposits and groundwater in the valley bottom are thus contained within a bedrock trough.

The valley-fill sediment (fig. 7) was laid down in glacial lakes and river plains during glacial retreat. Tunnel channels on the valley floor at the base of the glacier, and channels draining into the valley along the glacier margin on Ramapo Mountain and, to a lesser extent, the basalt ridges on the southeast side of the valley, carried meltwater that delivered sand and gravel into the lakes. The tunnel channels deposited lacustrine fans on the lake bottom at the glacier margin. As the margin retreated, the fan deposits were left behind as an irregular basal layer on the bedrock surface. Where the margin stabilized for several years the lacustrine fans aggraded to lake level and built outward into the lake as a delta. Small deltas and alluvial fans also formed along the valley wall where the ice-marginal channels entered the glacial lakes. The total thickness of glacial-lake deposits is as much as 200 feet.

There were two lakes in the main valley. The southwestern third of the valley (downvalley from ice margin M3 on fig.7) was occupied by the Totowa stage of glacial Lake Passaic. This lake was held in by a sediment dam that blocked the Passaic valley at Totowa, about 10 miles south of the Ramapo Valley. The level of this lake in the Ramapo Valley was 220-230 feet. Deltas and fans deposited from ice margins M1, M2, and M3 in this lake filled it completely, and fluvial topset beds at the top of the valley fill aggraded to a thickness of as much as 70 feet at ice margin M3. Most of the lake-bottom deposits in this lake are lower foreset and bottomset beds laid down as deltas prograded, and so are primarily silt and fine sand.

As the ice margin retreated from M3, the aggraded topset beds filling the narrow stretch of the valley south of M3 acted as a dam for a second, higher lake in the valley north of M3. This lake was initially at a level of 290-300 feet but by the time the margin had retreated to M6 erosion of the deposits downstream from M3 had lowered the lake level. Fans and deltas deposited in this lake are not as large as those deposited at and south of M3. Deltas are mostly restricted to the west valley wall, where they were deposited from ice-marginal channels descending Ramapo Mountain. Lacustrine fans on the valley bottom did not aggrade up into deltas. This drop in sediment volume is due to either (1) the ice margin retreating more rapidly north of M3, or (2) subglacial meltwater drainage partially diverting into the Masonicus Valley to the east of the main valley. The reduced influx of coarse sediment, and the widening of the valley, north of M3, allowed clayey, more continuous lake-bottom sediment to accumulate in this reach of the valley.

As ice withdrew northward into New York State the outlet for the second lake continued to lower as erosion south of M3 progressed. Clay and silt accumulated, raising the bed of the lake. Lowering of the outlet and filling of the lake eventually exposed the lake bottom. Meltwater coming down the valley from ice margins to the north deposited a river plain of sand and gravel on the exposed lake bottom. The plain continued downvalley as a narrow terrace south of M3. This river-plain deposit is less than 25 feet thick.

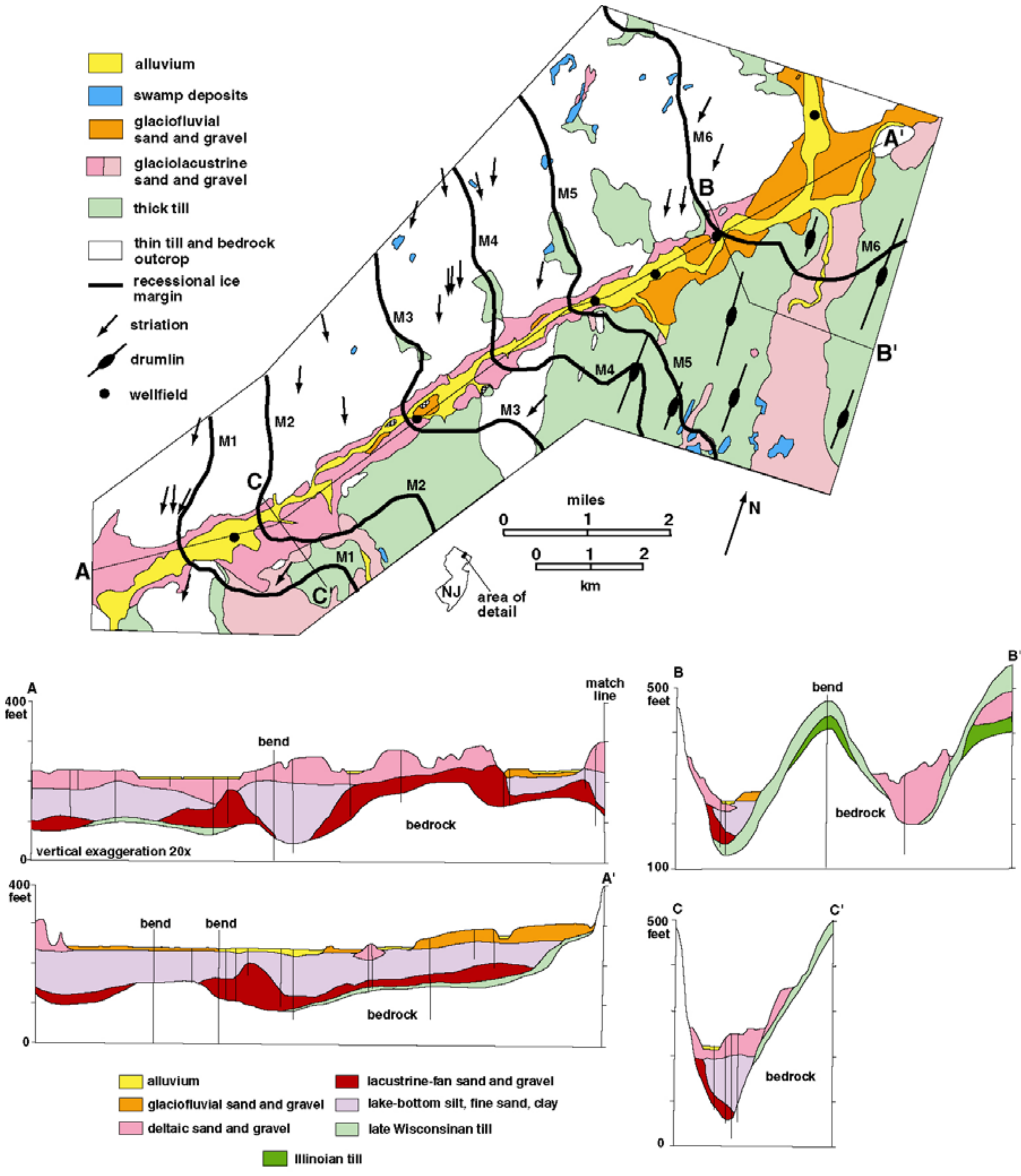

Figure 7. Map and sections of the Ramapo valley fill.

When ice withdrew from the Ramapo basin, sediment load and stream discharge declined. The modern Ramapo River began to establish its channel and cut a floodplain into the glacial valley fill. Return of forest cover with warming climate further reduced sediment influx to the river and stabilized the floodplain and channel. The floodplain deposits are generally less than 10 feet thick.

The valley-fill sediments left as the record of the above events are typical of a filled lake basin. They include an upper, unconfined sand and gravel aquifer, a middle confining or semi-confining unit, and a lower sand and gravel aquifer. The upper sand and gravel includes deltas and the river-plain deposit. Where the confining unit is absent, the upper aquifer extends to bedrock and may include lacustrine-fan deposits beneath deltas. The confining unit includes silt, clay, and fine sand laid down on lake bottoms and in the bottomset and lower foreset beds of deltas. In the Ramapo valley fill, the portion of the confining unit north of M5 is primarily a lake-bottom deposit; south of M5 it is primarily a bottomset and lower foreset deposit. Thus, the confining unit is more clayey north of M5 and is mostly silt and fine sand south of M5. Groundwater modeling indicates that the clayey portion of the unit acts as a true confining layer whereas the silt and fine sand do not (Hill and others, 1992). The lower sand and gravel occurs only below the confining unit and consists of lacustrine-fan deposits and possibly a little collapsed deltaic sand and gravel.

The thickness of these three layers can be mapped from records of wells and borings, and from surface exposures (fig. 8). Thick sections of upper aquifer mark stable ice margins where lacustrine fans aggraded up into deltas (chiefly south of M3), or where deltas built into the lake on the side of the valley at the mouths of ice-marginal channels (north of M3). In these settings the confining unit is thin or absent. Conversely, the confining unit is thickest between stable ice-margin positions (south of M3)and in the valley center, away from the ice-marginal channel mouths (north of M3). The lower aquifer is thick in a narrow, elongate string in the deepest part of the valley fill. This geometry marks the location of the subglacial tunnel channel that deposited the lacustrine fans comprising the lower aquifer.

Seepage from the Ramapo River is an important recharge mechanism. Measurements of stream flow and computer modeling of groundwater flow show that water is drawn into the valley fill from the river as it passes by wellfields (Vecchioli and Miller, 1973; Hill and others, 1992). Water is also lost from the river as it crosses more permeable parts of the valley fill, even in the absence of nearby pumpage (Hill and others, 1992). Where the confining unit is present between the river and wells pumping from the lower aquifer, seepage loss from the river may be shifted up or downstream to places where the confining unit thins or is absent. Seepage loss also occurs on tributary streams when they cross from till and bedrock uplands onto the deltaic sand and gravel along the valley walls, with some streams drying completely before they reach the main river.

Potential zones of seepage loss can be identified by mapping where the Ramapo River and tributary streams flow on valley fill where the confining unit is not present (fig.8). These spots include places where the river swings against the valley-side deltas and where the river cuts through the ice-contact zones of deltas that connect directly in the subsurface to lacustrine fans. Water entering the valley fill at these points may descend and then move laterally into the lower aquifer beneath the confining unit.

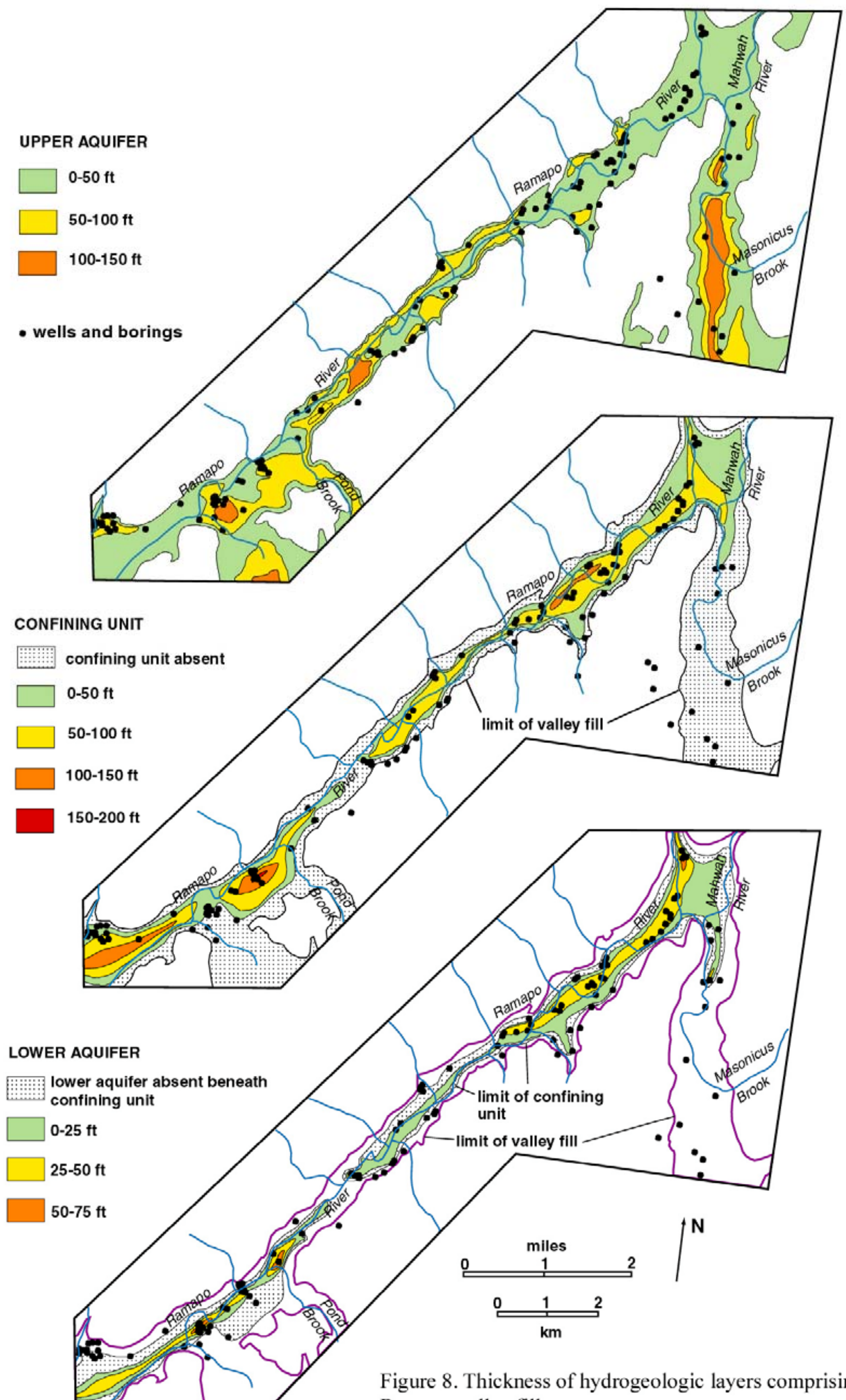

Figure 8. Thickness of hydrogeologic layers comprising the Ramapo valley fill.

CONCLUSIONS

The New Jersey Highlands were glaciated three times within the past two million years. A pre-Illinoian glaciation sometime between 2 Ma and 800 ka covered the entire Highlands. Deposits of this glaciation are deeply eroded and intensely weathered. An Illinoian glaciation at about 150 ka covered the northern half of the Highlands. Deposits of this glaciation are eroded from steep slopes but remain on gentle slopes, are only moderately weathered, and form thick fills in several valleys. The late Wisconsinan glaciation between 22 and 18 ka covered the Highlands at and north of the terminal moraine. Deposits of this glaciation are largely uneroded and only slightly weathered.

Glacial-lake and glacial-river deposits of late Wisconsinan and Illinoian age form locally productive valley-fill aquifers. Fluvial valley fills are unconfined sand and gravel deposited by glacial rivers in valleys that did not contain lakes. Unfilled lake basins were large glacial lakes that contain glaciolacustrine sand and gravel aquifers confined in places by fine-grained lake-bottom sediment. Filled lake basins were smaller glacial lakes that contain both a lower, confined sand and gravel aquifer and an upper, unconfined sand and gravel aquifer separated by a confining layer of fine-grained lake-bottom sediment. Stacked valley fills occur in glacial lakes along the terminal moraine and contain multiple layers of till, glaciolacustrine sand and gravel, and fine-grained lake-bottom sediment. The valley-fill deposits, and till on uplands between valleys, store and release groundwater to a greater degree that the underlying gneiss bedrock and so are an important source of stream baseflow.

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