Glacial Geology of New Jersey

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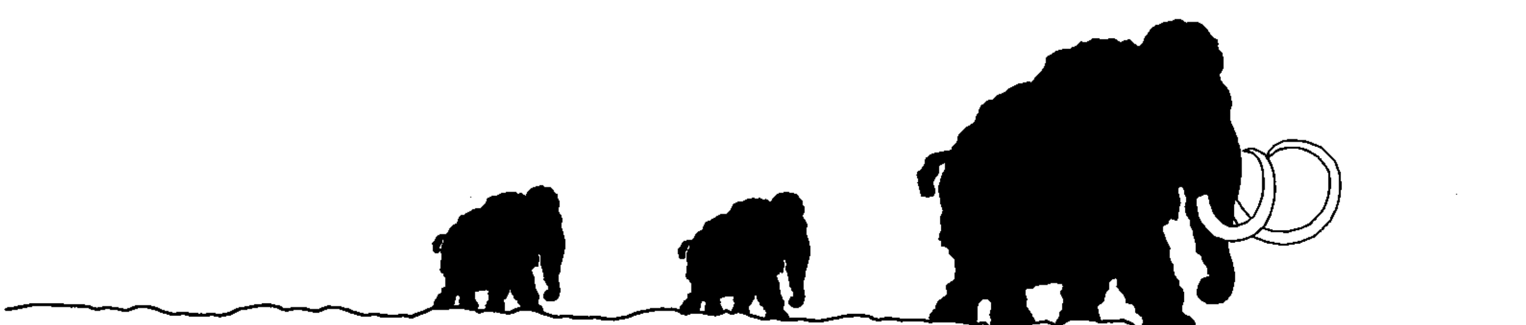

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CHAPTER 1

INVESTIGATIONS OF NEW JERSEY GLACIAL DEPOSITS, 1840-2000

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ABSTRACT

Continental glaciation was generally accepted by North American geologists by the late 1860s. Although New Jersey is at the southern limit of continental glaciation and might be considered important to the understanding of North American history, New Jersey's glacial deposits were not carefully investigated until the late 1870s. The New Jersey Geological Survey had mapped the southern limit of the terminal moraine by 1877. Over the next three years, the Survey identified recessional moraines and numerous glacial lakes. Mapping could not be completed, however, without an adequate topographic base. Topographic mapping was completed statewide between 1877 and 1887, and mapping of the glacial deposits was carried out between 1891 and 1912 as part of the U.S.G.S. - N.J.G.S. cooperative project for the creation of the 1910-1912 "Geologic Map of New Jersey". After the publication of the map, formal mapping of glacial deposits in New Jersey came to a virtual standstill. Topical studies of the New Jersey Pleistocene continued parallel with general understandings of Pleistocene geology elsewhere, and New Jersey is particularly known for geomorphic studies of the cycle of erosion and peneplain development and for the controversial Abbott Farm archeological site near Trenton, believed for a time to prove human habitation of North America in the Pleistocene. Particularly since the 1950s, ground water resources have been a driving force in the investigation of glacial deposits, and a substantial body of hydrogeologic work on the "valley fill aquifers" had developed by the late 1970's. By this time, the need for systematic updating of New Jersey's geologic mapping had become clear. Water resource bonds partially funded a second U.S.G.S. - N.J.G.S. mapping project between 1984 and 1990. Two summary maps of the glacial deposits have been prepared from this work, one using formal stratigraphic nomenclature, the other using a hydrostratigraphic approach.

INTRODUCTION

Whether or not continental glaciation had occurred was an important and widely popularized controversy from the 1830s until the 1860s, but the acceptance of glaciation does not rank among the most significant geologic advances of the century. Comparing the controversy over continental glaciation to those of modern memory, it might be considered equivalent to the controversy over extinctions by meteorite impact, but not to that over continental drift. The highly popularized mass extinctions by meteorite impacts are important, but did not have a pervasive impact on geologic thought. The idea of moving continents, on the other hand, affected the entire fabric of geologic thought. The two nineteenth century geologic advances that most pervasively changed the way natural scientists viewed the world were evolution and recognition of the extent of geologic time. Ice sheets covering much of the "civilized world" were important, particularly in understanding human history, but did not transform the way natural scientists viewed the world to nearly the extent that evolution and geologic time did.

While continental glaciation is commonly thought of as challenging the belief that unconsolidated surficial deposits of the northern parts of Europe and North America were left by Noah's flood of the Bible, most natural scientists had abandoned the idea of a single "Noah's flood" well before continental glaciation was proposed. Consistency with sacred history remained important, but strict Biblical literalism had fallen out of favor as a line of inquiry into natural history several decades earlier. In addition, geologists knew well before mid-latitude glaciation was proposed in a well-developed form in 1837 that the deposits were difficult or impossible to explain by flooding even if more than a single flood was envisioned. Flooding did not, for example, explain the restriction of the deposits to northern latitudes, the presence of house-sized "erratic" boulders many miles consistently southward from their source, and the lack of sorting of the deposits we now know as till.

The concept of formerly widespread glaciation first began developing in areas around the Alps beginning about 1820, and was presented in a well developed form for northern Europe by Louis Agassiz in 1837 (Flint, 1971). Earlier in the century, Charles Lyell and others had developed convincing interpretations attributing glacial sediments to icebergs carried by floodwaters. Whalers had reported that the icebergs in both Arctic and Antarctic seas were often loaded with stones and dirt. Melting of debris-laden icebergs explained the northerly distribution of the European deposits, the southward transport of large erratics, and the poor sorting. In addition, icebergs, unlike floodwaters, provided an effective means for carving drumlins and striating bedrock.

Both the glacial theory and the iceberg transport theory were developed primarily in Europe, but the idea of iceberg transport of erratics may have been independently conceived in North America in 1825 by Peter Dobson of Vernon, Connecticut. Based primarily on observations of till in the Vernon area, Dobson wrote, "I think we cannot account for these appearances unless we call in the aid of ice along with water and that ... [the boulders] have been worn by being suspended and carried in ice, over rocks and earth, under water." (Dobson, 1825).

Attribution of glacial features to icebergs became widely accepted prior to the development of evidence for continental glaciation, and abandonment of the iceberg transport theory was slow. Significant opposition to the idea of mid-latitude continental glaciation continued into the 1860s in England, the United States, and Germany, and individual geologists continued to oppose the idea even into the 1890s (Flint, 1971).

RECOGNITION OF GLACIATION IN NEW JERSEY

With New Jersey being at the southern limit of glaciation, one would expect the state to be a focal point of early and continuing work. Instead, much of the early work was done in New England following Louis Agassiz's move from Europe to Harvard University. The first concentrated work on New Jersey's glacial deposits was done in the 1870s, after the glacial theory had been widely accepted.

Henry Rogers' "Geology of New Jersey" (1840) is typical of early geologic work in concentrating on bedrock geology and older unconsolidated deposits (in this case those of the New Jersey coastal plain) almost to the exclusion of glacial deposits. Rogers refers to the glacial deposits of northern New Jersey and the surficial gravels of southern New Jersey as diluvium. Diluvium and drift were widely used terms in the 19th century. Diluvium, like "dilute", is from the Latin for washing away, and refers specifically to flood deposits. "Drift" originally referred to sediments carried by drifting icebergs, as discussed above. Later it came to be associated with a "drift period", but still did not indicate glacial deposition. The term came to refer to glacial deposits only after the general acceptance of glaciation.

Rogers' discussion of diluvium is primarily to note that it obstructed investigation of bedrock, hid areas likely to contain iron ore, and supplied water to wells. Because description of the diluvium is included in sections on the formations it overlies, and the diluvium is not discussed statewide in general terms, it is not clear from the 1840 report whether Rogers attributed all the diluvial deposits to a single flood. A single, brief episode of flooding would be reasonable, however, in view of Rogers' interpretation of other rock groups as resulting from extended periods of deposition by normal alluvial or oceanic processes terminated by catastrophic regional upheavals and consequent violent flooding.

With regard to the diluvium, coarse, heterogeneous gravel near the Flatkill, Sussex County is cited as "evidence of the short duration and violence of the action by which the miscellaneous debris from the adjacent rocks was hurled together." Further south, presumably simultaneous flooding was interpreted as sweeping along the Delaware, curving southward, removing several tens of feet of Tertiary sand and gravel overlying the greensand formations of the lower Delaware Valley and cutting away "a large portion of the greensand."

By the 1860s, glaciation and iceberg transport had become active subjects of inquiry in New Jersey. Dwight (1866) attributes an erratic and striations in

Englewood to glaciation. George H. Cook in his 1868 "Geology of New Jersey", uses the flood-related term "diluvium", but clearly realized that the surficial deposits were complex, and he saw the effects of current-driven flooding, quiet water deltaic deposition, and ice. Flood deposits are described as follows:

> "In some places banks of gravel or sand will be found deposited behind some protecting ridge of rock, just as would happen in a stream now. At Sand Hills, southwest of New Brunswick, at Jersey City, and other places, similar deposits of sand behind rocks are to be seen."

Lacustrine deposition is recognized in other passages:

> "In some localities, as at the Long Meadow in Warren County, Succasunny Plains in Morris County, and portions of the Passaic Valley, the fine, loamy soil, perfect freedom from boulders and coarse gravel, indicate that the soil has been deposited in the still waters of a lake or pond; and along the Delaware above the Water-Gap, and in the Ramapo Valley, ... the large and level-topped sand and gravel hills lead to the inference that these valleys were once filled with still water, and the streams which ran into them have deposited their deltas of sand in these banks which are now the smooth terraces of the valleys."

The report also recognizes a time of frigid climate and erosion by ice. The following description of the ice effects is not entirely specific and could refer to either icebergs or glaciers. Its index entry, however, "Ice, effects of moving" seems to refer to glaciers rather than icebergs.

> "The rounded surface of the rocks, in the Highland, the Paleozoic, and the Trap ridges, the regular and parallel scratches upon these surfaces, and the deep furrows worn in the softer rocks, all prove that some more rigid force than that of water has been in operation all over the country. ... it is only necessary to say that these effects, as well as the carrying of boulders, point to ice as the effective agent in producing them. Two skulls of the walrus, an animal living only in polar seas, has been found in the gravel near Long Branch. They indicate a period of cold more severe than any that now prevails. "

Ten years later, Cook's "Annual Report of the State Geologist for 1877" (Cook, 1877) unequivocally accepts glaciation, describing the drift as "undoubtedly ... produced by the action of a great glacier, which has covered the whole northern part of our country in recent geological times." The report includes a map tracing the terminal moraine across New Jersey at very nearly the location shown on current maps.

Sustained work on the glacial deposits continued over the next three years, and the "Annual Report of the State Geologist for 1880" (Cook, 1880) includes an extensive discussion recognizing recessional moraines, extinct lakes in most of the larger valleys, and the glacial origin of almost all of northern New Jersey's lakes. Several lakes which demonstrate glacial processes particularly well, including Budd Lake and Green Pond (now Green Lake) in Morris County, are described in some detail. Shorelines are traced narratively for two of the larger glacial lakes, Lake Passaic and Lake Pequest, and are shown on a map for Lake Passaic.

The terminal moraine is described in greater detail than in the 1877 report, but older glacial deposits to the south are not confirmed. Relatively fresh materials now attributed to the Illinoian are included in the terminal moraine and not specifically mentioned. Highly weathered materials to the south of the Illinoian deposits were noted as having many characteristics of glacial sediments and as being difficult to explain by either normal river processes or flooding. In the absence of convincing evidence of glacial origin, however, they were attributed to flooding related to the glaciation responsible for the terminal moraine.

While the general pattern of glacial retreat and lake formation outlined in the 1880 report were verified by later work, details of the depositional processes appear to be based on understandings which were developed outside of New Jersey and either did not prove appropriate to New Jersey or have not held up under subsequent work. In many of these instances, deposits now interpreted as formed in intimate contact with the melting ice were described in 1880 as formed at some distance from the retreating glacier. In three examples:

1) Deltas bordering New Jersey's lakes were described in numerous places as moraines eroded into the form of terraces. In one description, "The old glacial dams were not disturbed beyond a leveling of their surface and a sorting of the materials at the top." The emphasis on erosion rather than deposition may reflect contemporary work on erosional landforms bordering the Great Lakes.

2) Some stratified deposits not clearly related to lakes were interpreted as the result of torrential floods caused by rapid melting or breaching of dams. From terraces at relatively high elevations along Flat Brook, Sussex County, for example, the report infers a glacial river "more than a mile broad and at least 40 feet deep" which flowed into the Delaware. While flooding certainly did occur during glaciation, many of the deposits interpreted in the 1880 report as the result of flooding are now interpreted as deposited in lakes and ponds, some very small and temporary, in direct contact with the ice.

3) Deposits of Lake Hackensack and Lake Paramus are attributed to marine processes during the high sea level of the "Champlain epoch". High sea level features of the Champlain Seaway of New York, Vermont, and Canada resulted from isostatic subsidence of the crust under the weight of glacial ice. Isostatic subsidence in New Jersey is now known to have been far less than glacial sea level lowering, and the shoreline was seaward of its present location for the first several thousand years of glacial retreat.

It is not clear from the New Jersey Survey's annual reports who actually did the work between 1877 and 1880. The reports do not generally attribute credit other than to George Cook as State Geologist. Likely, however, it was done by John C. Smock, who is identified in the administrative portion of the 1877 report as working on the moraine and who continued an interest in the glacial deposits until at least the 1890s, when Salisbury (1891) credits him with recognizing older glacial deposits south of the terminal moraine.

After the acceptance of glaciation, studies of New Jersey's Pleistocene deposits paralleled those done elsewhere. Significant studies have been done on pollen, lake sediments, vertebrate paleontology, and local chronologies of ice advance and retreat, and New Jersey is particularly associated with historic developments in geomorphology and Pleistocene archaeology. As most of these issues are dealt with elsewhere in this volume the following comments refer only to geomorphology, archaeology, and vertebrate paleontology.

Geomorphology

From the 1880's through the 1930's, early North American work on landform development (Davis and Wood, 1890; Davis, 1890; Salisbury, 1898; Johnson, 1931) was carried out at Columbia University and elsewhere with considerable attention to New Jersey landforms. The Schooley and Somerville peneplains have their type areas here. Cycles of landform development beginning with uplift and ending in peneplanation proved overly schematic, and landforms are now more realistically seen as the result of wide ranges of constantly changing processes. The work on erosional processes, however, remains a starting point for all subsequent geomorphic theories.

Most of the landform work was not directly related to glaciation, but several of the major stream captures important to this history have been attributed to ice effects (see, for example, Campbell and Bascom, 1933; Lovegreen, 1974; Stanford, 1997; Witte, 1997).

Pleistocene Archaeology

Prior to 1872, there was general agreement that Indians were relatively recent immigrants from Europe or Asia, having arrived at most a few thousand years before the present. In that year, Charles Conrad Abbott noted similarities between Paleolithic tools from Europe and implements from gravel terraces overlooking the Delaware River south of Trenton (Abbott, 1872). As summarized by Kraft, 1993, Abbott had become convinced by 1876 that the gravels dated from the ice age and that the implements proved the antiquity of human occupation of North America (Abbott, 1876). In 1881, Abbott presented a comprehensive exposition of his findings which polarized the American archaeological and geological communities between those who supported his views (among them Frederick W. Putnam and Nathaniel Shaler of Harvard, and Henry C. Lewis of the Pennsylvania Geological Survey) and those who were antagonistic (chief among them William H. Holmes of the United States Natural History Museum).

Over the next several years the debate became increasingly rancorous. Numerous investigations confirmed the Abbott farm as one of the richest archeological sites in eastern North America, and made it one of the best known. Convincing proof of the antiquity of the site was not found, however, and an 1897 conference concluded that the implement-bearing deposits were not Pleistocene. At his death in 1919, Abbott was one of the few firm believers that human antiquity had been demonstrated in North America (Kraft, 1993).

The Abbott Farm site has been investigated repeatedly since then, most recently in the 1980s. It is now a National Historic Landmark recognized for repeated habitation from before 8,000 years ago (Middle Archaic) through at least 270 BC (Late Woodland). Its greater influence, however, may lie in the dampening the spirit of inquiry into human arrival in the New World. As phrased by Frank Roberts, Jr. (1940), of the Smithsonian Institution "The question of early man in America had become virtually taboo, and no anthropologist, or for that matter geologist or paleontologist, desirous of a successful career would tempt the fate of ostracism by intimating that he had discovered indications of a respectable antiquity for the Indian."

A respectable antiquity was finally demonstrated in 1926 when fluted Folsom projectile points were found together with extinct mammals. Although fluted points have been found throughout New Jersey and three Paleo-Indian sites are recognized in the state, none of these findings are associated with extinct mammals.

Pleistocene Mammals

Extinct mammals are fairly widespread in New Jersey, occurring in bogs in the glaciated areas, in a few of the gravels of southern New Jersey, and offshore on continental shelf areas exposed during the glacial sea level lowering (Parris, 1983). There are no world renowned localities comparable to the cave deposits of Virginia or the tar pits of California, but some of the peat bog specimens are outstanding.

Mastodons have been known from New Jersey specimens since 1824 (Dekay, 1824), and, based on teeth, were originally thought to be carnivorous. The mastodons are of all ages, including young, and are commonly uninjured. This is consistent with the interpretation that they perished after becoming mired in the soft ground. The mastodons of New Jersey are well represented in museums, with several currently on mounted display.

<u>Cervalces</u>, a large ruminant with some characteristics of an elk and some of a moose, is more closely associated than the mastodon with New Jersey. Only two reasonably complete skeletons have been found, both from Warren County bogs (Parris, 1983). <u>Cervalces</u> was at one point proposed at the state fossil.

MAPPING OF THE DEPOSITS – 1891 - 1912

Generalized geologic maps showing New Jersey date from the early 1800s, but accurate delineation of formations was not possible before the 1880s because adequate topographic base maps did not exist. As noted several times in Cook, 1880, accurate topographic surveys are particularly important to glacial geology because of the importance of meltwater drainage. The 1880 delineation of glacial Lake Passaic, for instance, did not identify the low point at which meltwater would overflow the lake basin south of the terminal moraine. The map of the lake is ambiguous, but breaks in the shoreline could indicate overflow at either Far Hills or Moggy Hollow. (Moggy Hollow is now known to be the actual outlet.) In part to allow geologic mapping to proceed, a "New Jersey State Topographic Atlas" was completed between 1877 and 1887. The maps were at a scale of 1:63,360 (one mile = 1 inch). At this time, New Jersey was at the forefront of topographic mapping, and was the first state to undertake and complete topographic mapping at this detail.

Systematic mapping of the glacial deposits was begun in the summer of 1891 as part of a cooperative project of the New Jersey and United States Geological Surveys. The project included both bedrock and surficial geology and culminated in the Geologic Map of New Jersey (Lewis and Kummel, 1910-1912). The glacial work was done primarily by Rollin D. Salisbury, Henry Kummel, Charles Emerson Peet, and George N. Knapp. Summary reports were published by the New Jersey Geological Survey (Salisbury, 1902; Salisbury and Knapp, 1917). Several detailed maps covering most of the area from the terminal moraine northward were published in the U.S.G.S. Folio Series and in New Jersey Geological Survey maps (figure 1). Statewide surficial geology is included on the 1910-1912 "Geologic Map of New Jersey" and its 1931 and 1950 revisions, but the delineation suffers from overgeneralization. The entire suite of ice-related deposits is lumped into four units: terminal moraine, recessional moraine, stratified drift, and early drift.

Mapping of New Jersey's glacial deposits came to a virtual standstill after publication of the 1910-1912 "Geologic Map of New Jersey". Elsewhere in the northeast, glacial mapping continued until somewhat later. The last of the major early northeastern glacial mapping projects to be completed was the glacial geologic map of Connecticut, completed by Richard Foster Flint between 1927 and 1930. This map is particularly important because it was based on a regional stagnation model of ice retreat contrasting with the staged retreat envisioned by most earlier workers. The question as to whether the glaciers retreated by regional cessation of ice movement and melting in place or by stepwise retreat of active ice remained unresolved until the 1970s in large part because the detailed, regional mapping studies necessary to resolve the issue were not being done.

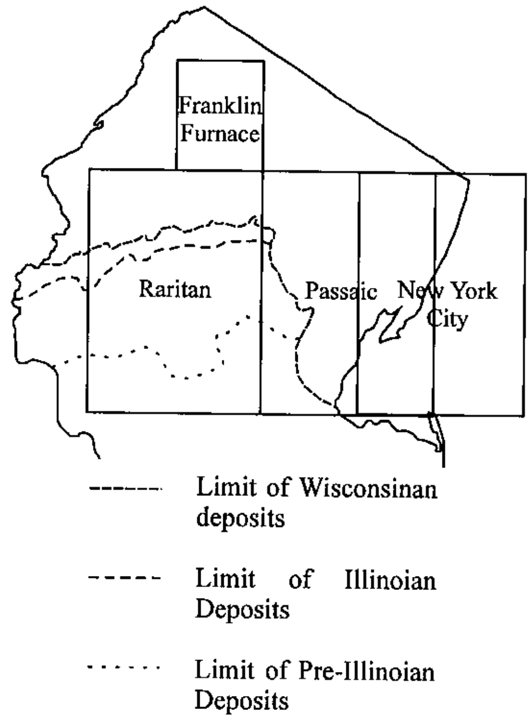

Figure 1. U.S. Geological Survey Geologic Atlas folios including glaciated areas of New Jersey: Franklin Furnace (Spencer and others, 1908), Raritan (Bayley and others, 1914), Passaic (Darton and others, 1908), New York City (Merrill and others, 1902).

WATER RESOURCES STUDIES

Water production from New Jersey's glacial deposits began early on, and was mentioned in early geologic reports. A number of studies were published concerning particular wells or well fields, and a great deal of informal work from the nineteenth century on is evident from correspondence on file at the New Jersey Geological Survey. A substantial body of work on the stratified drift and valley fill aquifers had developed by the 1960s, and work has continued through subsequent decades. Most of the published work is from cooperative investigations by the U.S. Geological Survey and the New Jersey Department of Environmental Protection, but academic institutions and consulting companies were active as well. Results were published as county reports and evaluations of particularly important aquifers.

The reports were prepared by geologists and hydrogeologists well versed in water resource issues rather than by glacial geologists. Origin of the aquifers was little explored other than being attributed to glacial, and in some cases preglacial, rivers and lakes. The work initially focused on horizontal and vertical delineation of the aquifers from lithologic logs and on potential yields based on pumping records. Beginning in the 1960's, mathematical modeling has been widely used together with the logs and pumping data for predictive evaluations.

While the water resources studies were satisfactory for delineation and evaluation of the central portions of the larger and more extensively drilled valleys (as, for instance, the upper Passaic River valley and upper Long Valley of Morris County), the reliance on well logs and pump test data almost to the exclusion of surface geologic mapping and depositional processes in glacial environments limited the detail of the interpretations, particularly in smaller valleys and less explored areas. Even in the major valleys, interconnections between aquifers and between well fields remained unclear in many instances due to absence of wells in critical locations.

MAPPING, INTEGRATION OF GEOLOGIC AND WATER STUDIES - 1980-PRESENT

In response to a growing need for up-to-date geologic information on both bedrock and surficial geology for water supply and environmental work, the New Jersey Department of Environmental Protection initiated a cooperative program in which the New Jersey and United States Geological Surveys completed the field work for new statewide surficial and bedrock geologic maps between 1984 and 1990. The mapping effort was partially funded by New Jersey Water Resources Bond funds, and was specifically directed towards ground water management.

In addition to a better understanding of glaciers, an abundance of resources not specifically related to glaciers had become available in the 70 years since the 1891-1912 project. These include 1:24,000-scale topographic quadrangle maps, aerial photography, abundant well records, and surface geophysical methods.

1:24,000-scale Topographic Maps

Depositional systems in New Jersey's glacial environments typically had extents of several square miles. Many of the glacial lakes were on the order of one square mile. At the mile to an inch scale of the 1880's New Jersey State Atlas, a square mile lake would cover the area of a large postage stamp. This square inch of map might include deltas, lake-bottom sediments, till, bare rock, post-glacial terraces, and alluvial deposits. The map would necessarily be overcrowded. At 1:24,000, on the other hand, this same area would cover a square just over 2-1/2 inches on a side. This scale is usually entirely adequate to show field mapping results.

An additional shortcoming of the 1880's maps is their somewhat generalized topographic contouring. In some areas it is difficult to judge the elevations of delta tops and divides with an accuracy of plus or minus 20 feet. In New Jersey's glacial lakes, several possible outlets are commonly within 20 to 30 feet of one another. Interpretation of lake history involves more than simple topographic reasoning, and can not be done directly from 1:24,000-scale maps any more than from mile-to-an-inch maps or aerial photos, but the additional topographic detail is far superior to the larger scale for field work as well as for presentation of results.

Aerial Photography

Mapping of Wisconsinan glacial deposits differs from bedrock mapping in that many of the depositional and erosional landforms still exist and are large enough to be seen on stereoscopic aerial photo pairs. Statewide aerial photography first became available in the 1930s. As with detailed topographic maps, aerial photography is enormously useful to effective field work.

Well Records

Lake deposits make up the bulk of New Jersey's glacial sediments. The lake deposits commonly form a vertical sequence including, from the bottom up, some or all of the following: 1) scoured bedrock, 2) till, 3) sand and gravel deposited at or near the retreating ice margin, 4) fine sand and silt deposited in open water further from the ice, 5) deltas at the ice margin, at meltwater streams entering the lake, and, in some cases prograding across the lake as it filled, and 6) post-glacial alluvium or peat.

A limited number of water well logs was available for the 1891-1912 mapping. These logs were often of very high quality and precisely described lithologic sequences. There were not enough of them, however, for confident generalization and the drawing of cross sections. Because of the scarcity of subsurface information, the early maps show only the deposits exposed at the surface.

After 1947, drillers were required to send well logs to the Department of Environmental Protection. Tens of thousands of well logs of varying quality are now on file. Many are useful only for estimating the depth to bedrock. A smaller number, but still in the thousands, are detailed and accurate, and allow confidence in drawing cross sections of the glacial deposits. Based on this abundance of data, it is possible to use map units which represent lithologies at depth as well as simply the material at the ground surface.

Surface Geophysical Methods

Well records are complete and invaluable in some areas and wholly inadequate in others. Poor well records may result from limited drilling in highly urbanized areas served by surface water, limited drilling in rural areas, or logging practices of particular drillers. In the recent mapping initiative, well records were inadequate, for example, to provide necessary information for water resources evaluation in the Great Swamp area of the Lake Passaic basin, the Millburn Gap area between the Lake Passaic basin and buried valleys to the east, along the Ramapo Valley to the south of aquifers in the Mahwah area, and in the Germany Flats area of rural Sussex County. In each of these areas, surface geophysical methods were used together with existing data and limited test drilling to provide data at a fraction of the cost of drilling alone. Seismic, electrical, and gravity methods were used.

Surficial geologic map coverage has been completed for nearly all of northern New Jersey at a scale of 1:24,000. In addition to surficial deposits, the maps include bedrock surface elevation data and contours, stratigraphic sections, and quadrangle-specific summaries of glacial and post-glacial history. Many of these maps have been published. The rest are available for inspection at the New Jersey Geological Survey.

Together with the 1:24,000-scale maps, the data have been compiled at 1:100,000 for a formal stratigraphic map using formation nomenclature and a separate glacial sediment map emphasizing hydrogeologic properties. As presented in the following papers (Stanford, Witte, this volume), the new geologic understanding builds on the earlier work and qualitatively advances the work in providing a 3-dimensional, lithologic framework for understanding the geology of the glacial deposits and for management of environmental issues and water resources.

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CHAPTER II

OVERVIEW OF THE GLACIAL GEOLOGY OF NEW JERSEY

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ABSTRACT

At least 3 glaciers have entered New Jersey within the past 2 Ma. Deposits of the earliest glaciation are intensely weathered, deeply eroded, and correlate to magnetically reversed drift in central and eastern Pennsylvania. They are of either late Pliocene (the pre-Illinoian K glaciation of the mid-continent, 2 Ma) or early Pleistocene (pre-Illinoian F or G, 800 to 900 ka) age. Deposits of the intermediate glaciation are moderately weathered and eroded. They correlate to till in New England that is probably of Illinoian age (150 ka). Deposits of the most-recent glaciation are only slightly weathered and generally uneroded. This glaciation, the late Wisconsinan, is dated by radiocarbon. Late Wisconsinan ice reached its southernmost position no earlier than 21 ka and had retreated north of New Jersey by about 18 ka.

INTRODUCTION

Over the past 20 years, measurements of the ratio of ^{18}O to ^{16}O in fossil marine shells, a ratio that is sensitive to the volume of glacial ice, indicate that Earth's northern hemisphere has had sizable glaciers for the last 2.5 Ma (Raymo, 1992). During glaciations the lighter ^{16}O isotope is preferentially evaporated from the ocean and stored in glacial ice, thereby enriching the ocean in ^{18}O. The ^{18}O to ^{16}O ratio in foraminferal shells, which reflects that of the seawater in which they reside, thus increases during glaciations, providing a continous record of ice volume (fig. I-1). About nine times over the past 2 Ma glaciers spreading from centers in northern Quebec and Nunavat (Northwest Territories) in Canada have grown large enough to enter the United States. These glaciers, known as the Laurentide ice sheets, are the world's largest. Their gradual growth and relatively rapid decay occur on a 100 ka cycle that shows up prominently as a sawtooth pattern on the oxygen-isotope record (ruled periods on fig. II-1).

Repeated Laurentide glaciations have occurred in the middle and late Pleistocene since about 800 ka and once in the late Pliocene at about 2 Ma (fig. II-1). These glaciations are characterized by a period of gradual buildup over 80-100 ka followed by rapid and nearly complete melting from maximum extent within 10-20 ka. The frequency of growth and decay is strongly correlated to variations in insolation in the northern hemisphere caused by changes in the ellipticity of Earth's orbit (eccentricity), in the tilt of Earth's rotation axis relative to the orbital plane (obliquity) and in the position of Earth on the elliptical orbit at equinoxes and solstices (precession). These orbital parameters (the Milankovitch cycles) do not by themselves provide sufficient insolation change to account for the 100 ka cycle of Laurentide glaciation. Rather they act together with Earth's climate system, in ways that are not fully understood, to pace glaciation. Glacial deposits are the direct record of glacier behavior and so, in addition to their economic, water-resource, and engineering importance, contain valuable information on climate.

At least three of these Laurentide glaciations extended into New Jersey and reached their southern limits here (fig II-2). These glaciers left a variety of sediments (known collectively as "drift") laid down in ice-contact, lacustrine, fluvial, and subglacial environments (fig. II-3). The drift is as much as 400 feet thick (fig. II-4) and forms an array of depositional landforms. Much of the glacial landscape consists of erosional forms carved into bedrock, with only a thin, patchy till cover (fig. II-3). The age of the most recent glaciation is known from radiocarbon dating; the age of the two earlier ones is not precisely known but can be estimated by correlating their deposits to adjacent glacial,

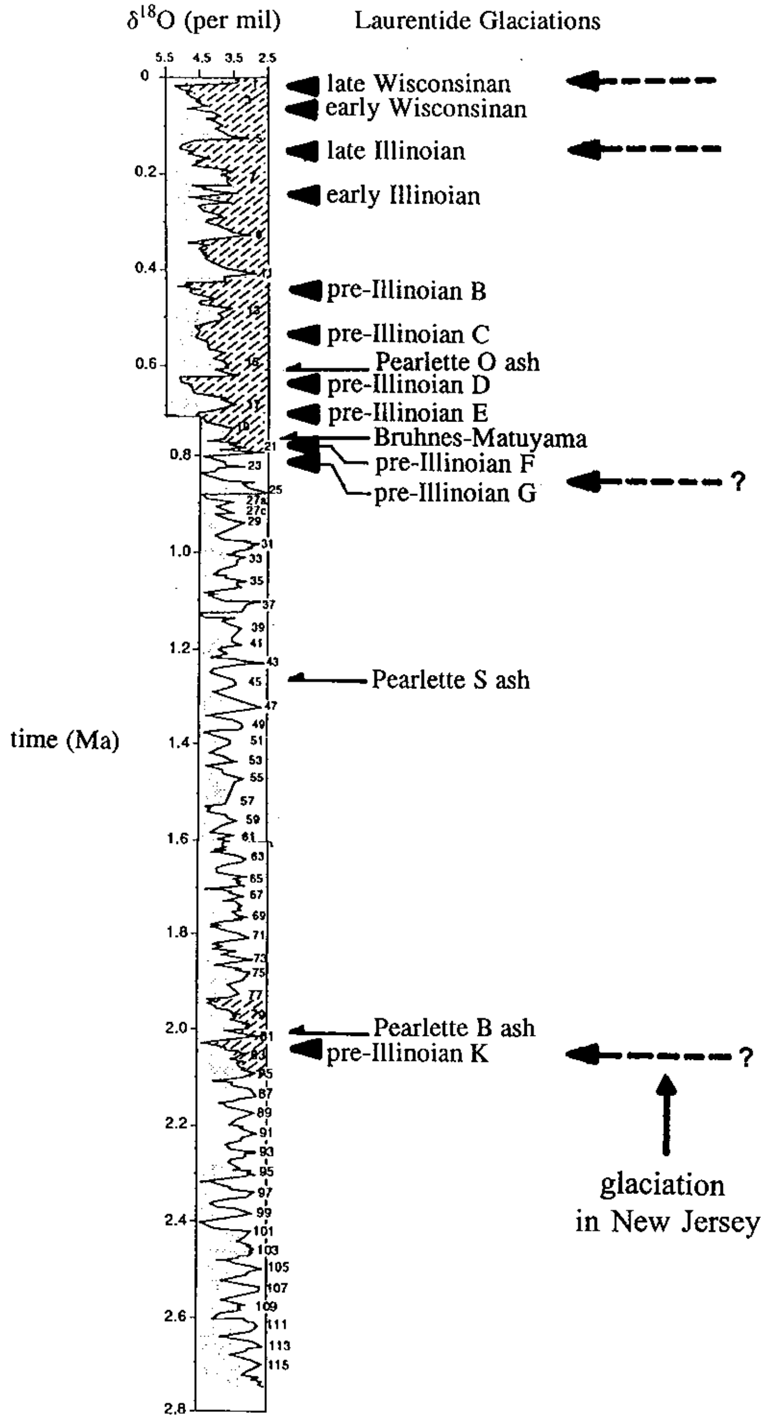


Figure II-1. Oxygen-isotope record from benthic forams at DSDP site 607 in the North Atlantic and Laurentide glaciations. Ruling indicates 100 ka periodicity possibly marking Laurentide glaciation. Oxygen-isotope curve compiled from Chappell and Shackleton (1986), Ruddiman and others (1986) and Raymo and others (1989). Laurentide glaciations and ash stratigraphy from Richmond and Fullerton (1986). Dashed arrows on right indicate glaciers that reached New Jersey.

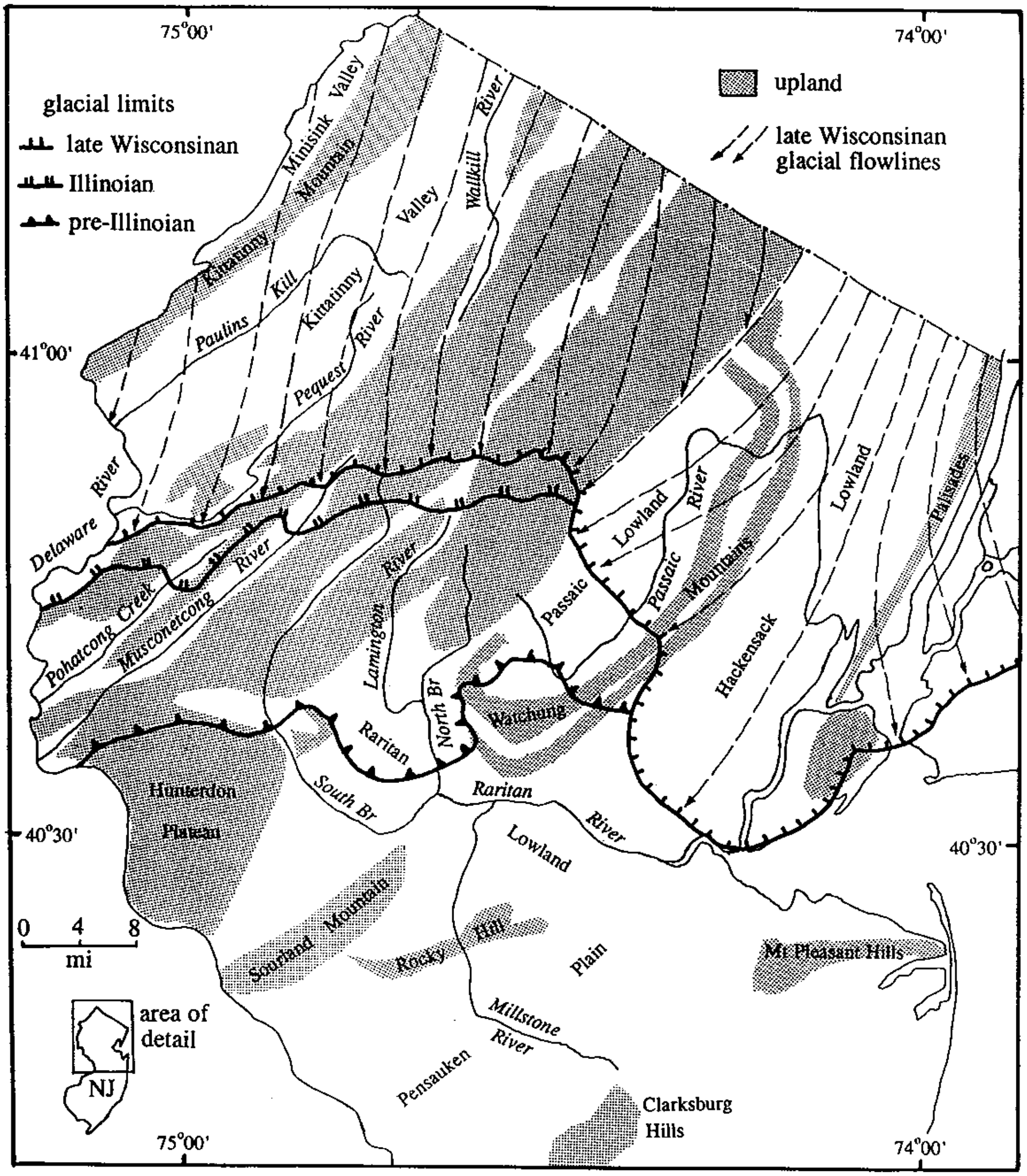


Figure II-2. Major physiographic features of northern New Jersey, with glacial limits and flowlines of the late Wisconsinan glacial advance. Flowlines are based on drumlin axes, erratic dispersion in till, and several hundred ridgetop and upland striations (Stone and others, in press). In the eastern half of the state the late Wisconsinan glacier has eroded and covered the earlier glacial deposits.

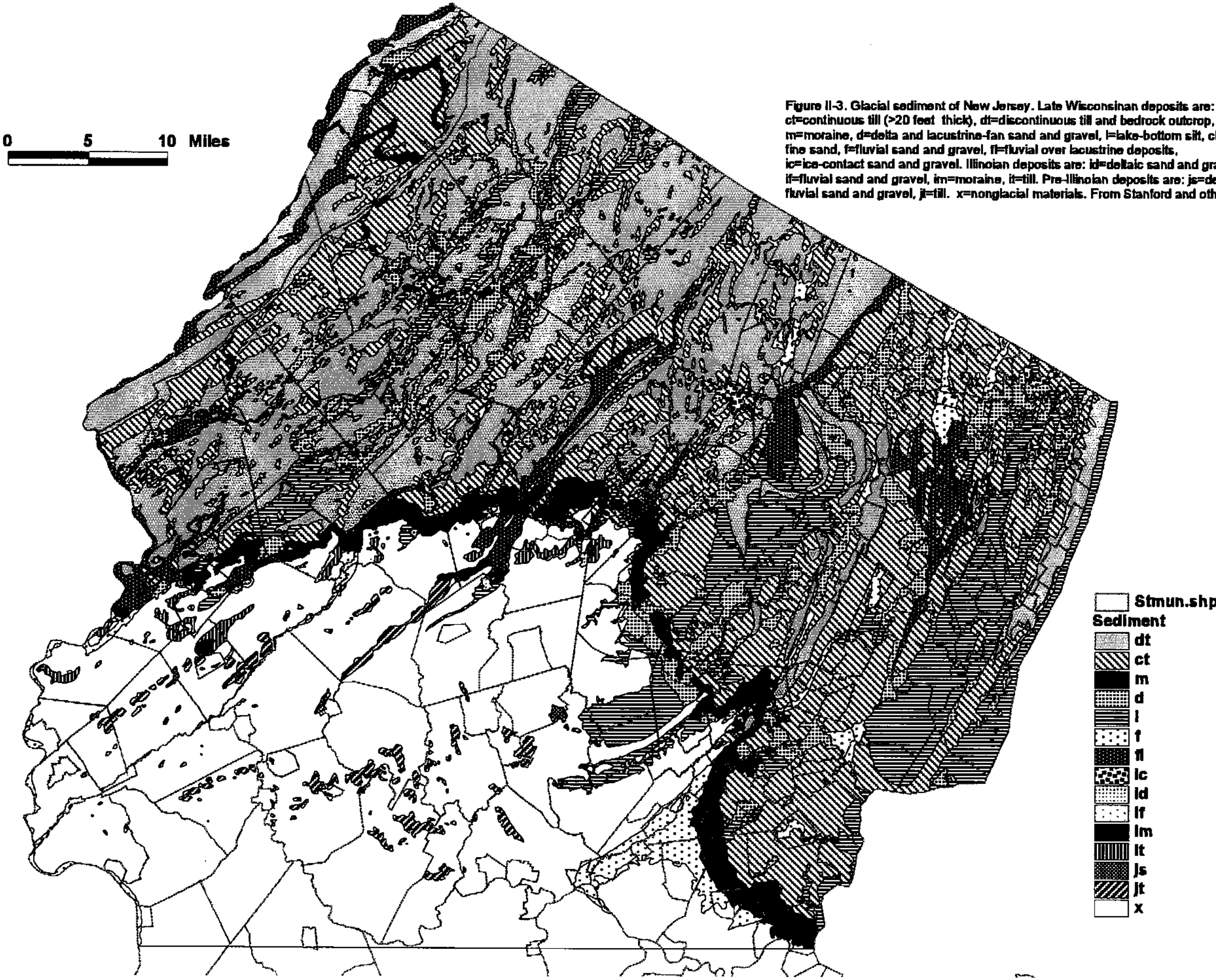
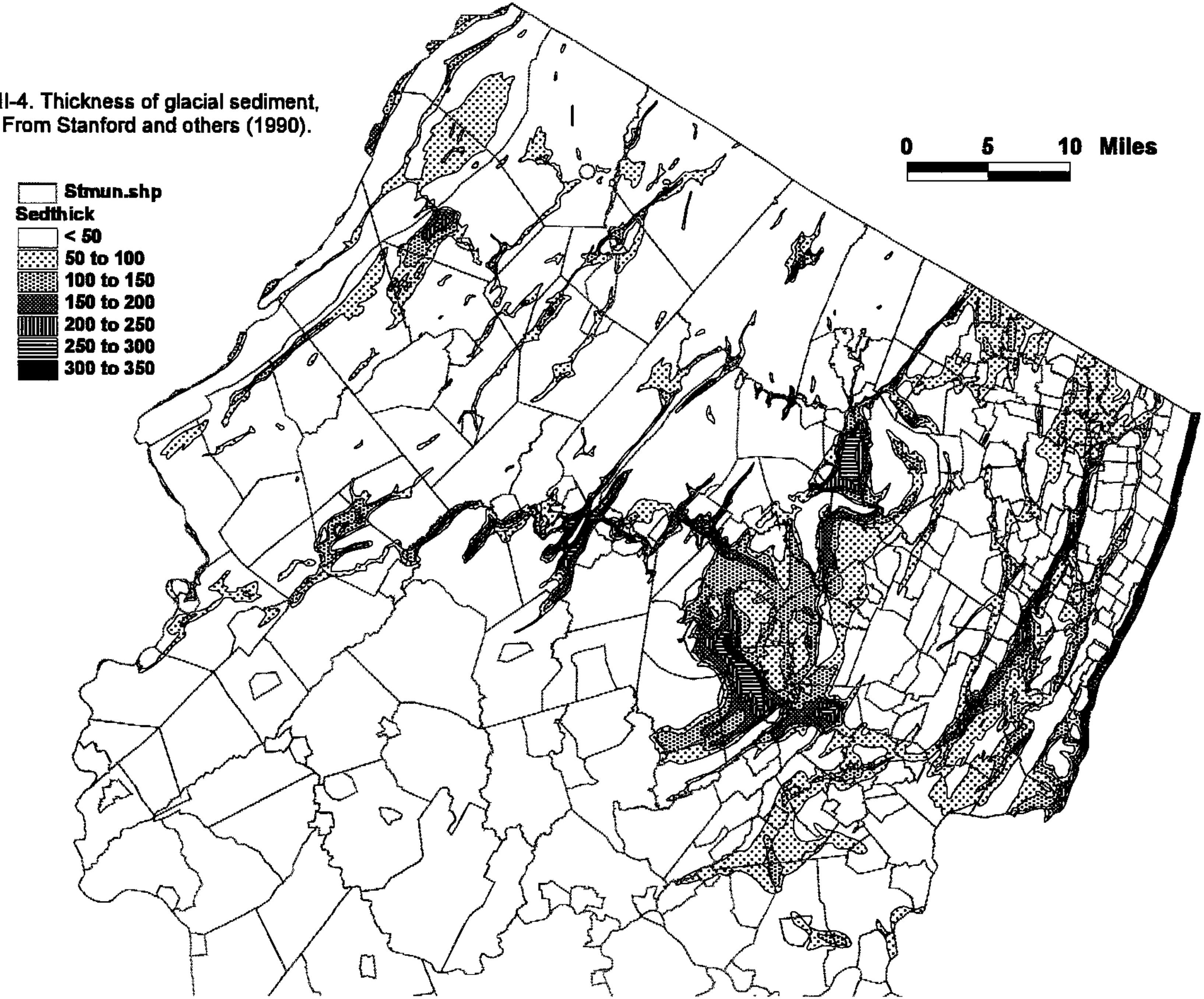


Figure II-3. Glacial sediment of New Jersey. Late Wisconsinan deposits are: ct=continuous till (>20 feet thick), dt=discontinuous till and bedrock outcrop, m=moraine, d=delta and lacustrine-fan sand and gravel, l=lake-bottom silt, clay, fine sand, f=fluvial sand and gravel, fl=fluvial over lacustrine deposits, ic=ice-contact sand and gravel. Illinoian deposits are: id=deltaic sand and gravel, if=fluvial sand and gravel, im=moraine, it=till. Pre-Illinoian deposits are: js=deltaic and fluvial sand and gravel, jt=till. x=nonglacial materials. From Stanford and others (1990).

Figure II-4. Thickness of glacial sediment, in feet. From Stanford and others (1990).



fluvial, and marine sediments and by comparing their degree of weathering and erosion.

This chapter will describe the deposits of these three glaciations, their effect on the landscape, and the evidence for their age. The material in this chapter is, in part, modified from the text for a Cook College field course on the hydrogeology of glacial deposits in New Jersey (Stanford and Ashley, 1992) and from the guidebook for the 1997 Friends of the Pleistocene field conference (Stanford and Witte, 1997).

PRE-ILLINOIAN DEPOSITS

The oldest glacial deposits, formerly termed the Jerseyan (Bayley and others, 1914) but now referred to as pre-Illinoian (Richmond and Fullerton, 1986), are deeply weathered and eroded. They were deposited by a late Pliocene or early Pleistocene glaciation (fig. II-1). They consist of till and a few small glaciofluvial and glaciolacustrine deposits (collectively designated as the Port Murray Formation in Stone and others [in press]) that are preserved in thin (<20 feet thick), patchy, deeply weathered remnants on flattish interfluve areas, chiefly on carbonate rock in the valleys of the Highlands and on shale in the lowlands of the Newark Basin. There are also a few patches of till on flats or saddles on gneiss and basalt uplands. In these landscape positions, the deposits have been protected from erosion (fig. II-5). Extensive erosion has removed the deposits from most of the area they formerly covered. Streams have cut valleys from 80 to 200 feet deep into bedrock below the former pre-Illinoian land surface, as evidenced by the absence of pre-Illinoian deposits on flats within these valleys.

In the surviving deposits, carbonate clasts in the till are completely decomposed to depths in excess of 15 feet; gneiss, sandstone, mudstone, and shale clasts are either completely decomposed or, if cobble-sized, have weathering rinds greater than 0.5 inches thick surrounding a partially weathered core; quartzite and chert typically have thin orange oxidation rinds and may be easily broken with a hammer but are not decomposed. In a few very well-drained locations, typically high-standing stratified deposits, clast weathering is less intense, and overlaps the degree of weathering observed in some poorly-drained Illinoian deposits. Soils are truncated but some remnants show red clayey Bt horizons as much as 6-8 feet thick.

Although no morainic landforms remain, the remnant till patches, and scattered erratic clasts, define a sharp outer margin for this drift. The margin is lobate and fitted to the location of ridges and lowlands, especially in the Raritan lowland, where the margin wraps around Cushetunk Mountain and around the west ends of the Watchung ridges, and extends southward about 8 miles in the broad intervening valley (fig. II-2). To the east it is overprinted by the late Wisconsinan moraine near Summit and has not been identified in the subsurface beneath late Wisconsinan deposits.

One possible pre-Illinoian glacial-lake basin can be inferred from the surviving deposits. A high-standing knoll of sand and gravel with a maximum elevation of about 460 feet near Bernardsville may be the remnant of a fan-delta deposited in a lake ("Lake Watchung" on fig. II-10) dammed between the pre-Illinoian ice margin and Second Watchung Mountain in the Passaic lowland. This lake may have been controlled by a spillway 4 miles to the south of Bernardsville at an elevation of about 430-440 feet across a gap in Second Mountain.

The age of the pre-Illinoian drift is uncertain but it likely correlates to magnetically reversed drift in central and eastern Pennsylvania (Gardner and others, 1994; Sasowsky, 1994), indicating a pre-788 ka age. Given the similarity of its weathering characteristics, topographic position, and erosional preservation to that of the Pensauken Formation, a preglacial fluvial deposit in central and southern New Jersey of Pliocene age (Stanford, 1993a), it may be correlative to the ash-dated late Pliocene pre-Illinoian K till of the midcontinent (fig.II-1) (Richmond and Fullerton, 1986). The potential for a late Pliocene age is also suggested by the marine oxygen isotope record. The sawtooth 100 ka periodicity in the record that is thought to reflect the climate inertia imposed by the Laurentide ice sheet occurs only in the last 800 ka (fig. II-1). A periodicity of 40 ka, which may not be sufficient for extreme Laurentide glaciation, dominates before 800 ka. The exception is a 100 ka periodicity leading up to stages 82 and 78 at 2.03 and 1.94 Ma respectively (Raymo and others, 1989), which corresponds to the 2.01-2.14 Ma range for the

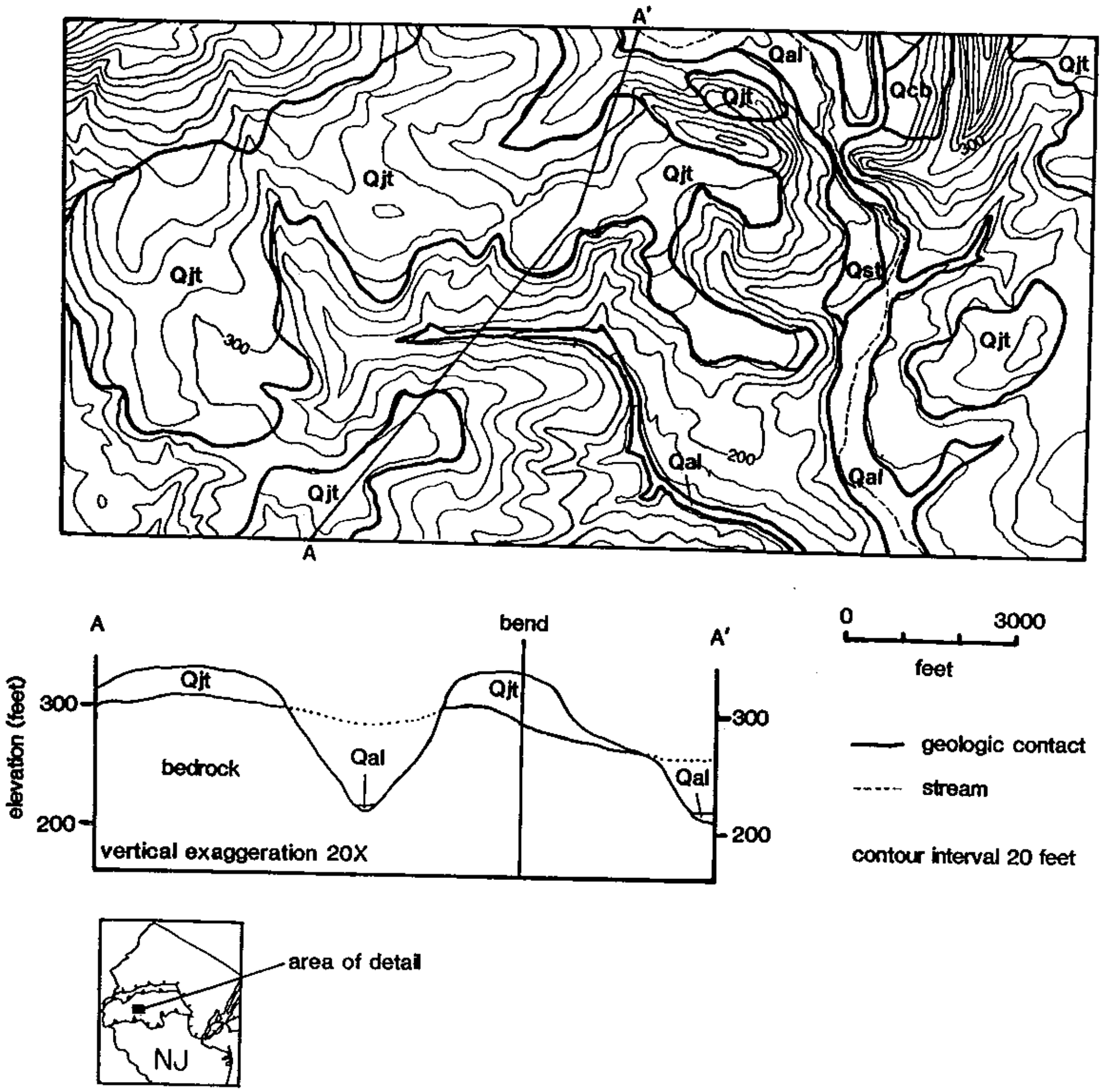


Figure II-5. Map and section of typical surficial deposits within the area of pre-Illinoian glaciation near Oldwick, New Jersey. Units are: Qjt=pre-Illinoian (formerly Jerseyan) till, Qst=nonglacial stream terrace deposits of late Pleistocene age, Qal=alluvium, Qcb=basalt colluvium. Unlabelled areas are bedrock, weathered bedrock, and thin, discontinuous colluvium. Dotted lines on section AA' mark the former land surface in pre-Illinoian time. Topography from U. S. G. S. Califon quadrangle. Note that pre-Illinoian deposits occur chiefly as thin erosional remnants on gently-sloping uplands.

age of K till in the midcontinent. If the pre-Illinoian till in New Jersey is magnetically reversed like those in Pennsylvania, then it correlates either with the pre-Illinoian F or G Laurentide glaciations between 900 and 790 ka or to the K glaciation at just after 2.0 Ma. Given the much closer ties of the pre-Illinoian till to the Pliocene Pensauken Formation than to the Illinoian drift, perhaps the 2.0 Ma age is more likely.

Another, more direct, bit of evidence suggesting a Pliocene age is pollen from a depth of 46-60 feet in a 60-foot core recovered by Harmon (1968) from Budd Lake, which occupies the headwater area of a small north-draining preglacial valley on gneiss bedrock in the central Highlands (fig. II-10). The lake is dammed on its north end by the late Wisconsinan terminal moraine, which overlies late Wisconsinan and Illinoian? lacustrine deposits that completely fill the preglacial valley (Stanford and others, 1996). However, the part of the lake in which the core was taken is outside the late Wisconsinan and, possibly, the Illinoian limit, but within the limit of the pre-Illinoian glaciation. Out of a total of 480 grains counted at 7 depths in the 46-60 foot interval, 64% were pre-Pleistocene exotic taxa including *Ephedra, Platycarya, Pterocarya, Podocarpus, Dissacites*, and *Phyllocladus*, and others not identified, in addition to a dominantly pine (32%), spruce (6%), birch (3-4%), alder (3%), and oak (10%) assemblage. These samples are from finely laminated gray clay that is sandy below 57 feet and includes a few pebbles. Above 45 feet the core shows the typical glacial to postglacial pollen sequence, with a date of 22,870±720 (I-2845) (see late Wisconsinan section below) at a depth of 35 feet. If the exotic pollen are in place, then they would imply a Pliocene glacial event because lake clays at the core location require glacial damming of the valley. Even if they are reworked from earlier deposits, as suggested by Harmon (1968), their abundance and preservation indicates storage in a non-oxidizing environment over a long period on this upland surface, which is most easily accomplished in a glacial lake or bog.

The pre-Illinoian deposits rarely exceed 20 feet in thickness because they are so extensively eroded. Owing to their deep and intense weathering they are generally not usable as sources of sand and gravel, although the pre-Illinoian till was formerly mined for brickmaking due to its high clay content.

ILLINOIAN DEPOSITS

The next glaciation is correlated with the Illinoian deposits of the midwestern United States. The Illinoian glaciation reached its maximum extent approximately 150 ka (fig. II-1), based on dating of the marine oxygen isotope record. In New Jersey, Illinoian deposits crop out in a narrow belt south of the late Wisconsinan limit, west of Morris Plains (fig. II-2). Illinoian sediment also occurs beneath late Wisconsinan deposits in the vicinity of the terminal moraine in the Delaware, Pequest, Musconetcong, and Rockaway valleys, and in the Passaic lowland between Morris Plains and Summit. Illinoian deposits have not been identified in the subsurface east of the Summit area, although a red, weathered, silty till (the Bergen Till of Stone and others [in press]) that cores some drumlins in northeastern New Jersey near the New York state line may be of Illinoian age. Illinoian till also occurs on the south shore of western Long Island, so Illinoian drift was almost certainly present in the Hackensack lowland before being eroded by the late Wisconsinan glacier.

Illinoian deposits, unlike those of pre-Illinoian age, lie within modern valleys (fig. II-6) and are not deeply weathered. Locally, soils developed in Illinoian deposits show red B horizons but generally the soils are stripped or truncated and red Bt horizons are rare unless preserved by burial under colluvium. Clast weathering varies considerably with site drainage. Poorly drained sites show nearly complete decomposition of gneiss, mudstone, and sandstone pebbles, and thick rinds on cobbles and boulders, which may have weathered and fractured cores. Moderately drained sites show rinds generally less than 0.25 inches thick on the same rock types, although some pebbles may be completely decomposed. Well-drained sites, typically on stratified deposits, show rinds less than 0.15 inches thick. At all sites, carbonate clasts are fully decomposed to depths of 10 feet or more, and quartzite and chert clasts are unweathered. Although clast weathering is variable, Illinoian deposits can always be distinguished from late

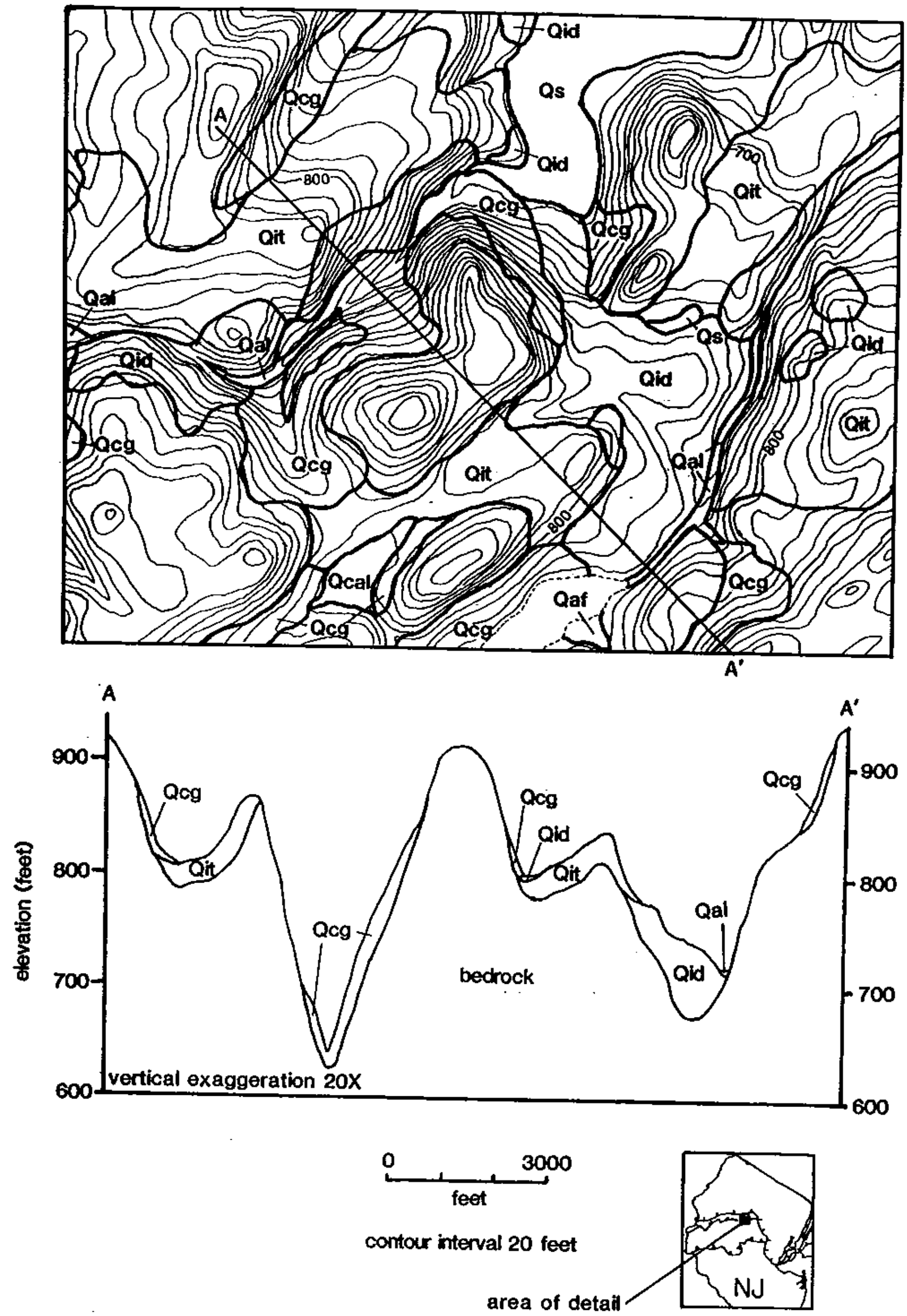


Figure II-6. Map and section of typical surficial deposits within the outcrop belt of Illinoian deposits near Dover, New Jersey. Units are: Qit=Illinoian till, Qid=Illinoian delta deposits, Qcg=gneiss colluvium, Qal=alluvium, Qcal=colluvium and alluvium, undivided, Qaf=alluvial fan deposits, Qs=swamp deposits. Unlabelled areas are weathered bedrock and thin, discontinuous colluvium. Topography from U. S. G. S. Mendham quadrangle. Note that Illinoian till is generally continuous on gentle to moderate slopes and that Illinoian deltaic deposits have subdued constructional morphology. Well-developed aprons of colluvium occur along the bases of steep slopes.

Wisconsinan deposits by the presence of some unweathered gneiss, mudstone, and sandstone clasts in the late Wisconsinan deposits, even at their southernmost extent, where they incorporate much weathered material. In Illinoian deposits these clasts always have at least a thin weathering rind. Distinguishing deeply weathered Illinoian deposits from pre-Illinoian deposits is not always possible on the basis of weathering intensity. Map-pattern criteria such as distribution and continuity on slopes are often more decisive.

On uplands within the Illinoian outcrop belt, till is more widespread than in pre-Illinoian terrain. It mantles gentle upland slopes but is stripped off of moderate to steep slopes, where the surficial mantle is chiefly fractured bedrock rubble and colluvium.

In valleys within the outcrop belt, Illinoian deposits form recognizable fluvial-plain, delta, and moraine landforms and are as much as 150 feet thick. Illinoian valley-fill deposits include three types: (1) glaciolacustrine deposits in the Passaic, Rockaway, Lamington, and Pequest basins, where glacial lakes formed in valleys that drained toward the Illinoian ice margin, (2) glaciofluvial deposits in the Musconetcong, Raritan, Pohatcong, and Delaware valleys, where glacial streams flowed down valleys that drained away from the Illinoian ice margin, and (3) morainal deposits (chiefly till) in the Lamington and Pohatcong valleys. The aquifer properties of these valley fills are discussed in chapter IV. Because they are generally not intensely weathered, especially where carbonate-rock content is low, Illinoian deposits have been mined at several locations for sand and gravel.

This drift is not directly dated but an Illinoian age is assigned based on the local presence of red Sangamon-like soil on the drift, and by correlation of the Bergen Till in northeastern New Jersey to the lower drumlin till of New England, which, in turn, is correlated to the lower till at Sankaty Head on Nantucket, which is overlain by a Sangamon marine sand dated by U-Th and AA on shells (Stone and Borns, 1986). An early Wisconsinan age is considered untenable based on the absence of sufficient ice volume, as inferred from the amplitude of the oxygen-isotope record, to allow glaciation this far south (Ridge and others, 1990). Salisbury (1902) and Ridge (1983) also suggested that this drift was partly formed during late Wisconsinan advance beyond the terminal moraine by incorporation of previously weathered surficial materials, but the sharp geomorphic boundary between the terminal moraine and patchy till-colluvium-rock rubble to the south, the distinct difference in weathering of the freshest clasts between the late Wisconsinan and Illinoian drifts, and the stacked tills and lacustrine sediment packages in the valleys beneath the terminal moraine all indicate separate glaciations. Non-morainic late Wisconsinan till does extend slightly beyond the southern margin of the moraine in a few places, but this till is uncolluviated, contains fresh gneiss clasts, and is readily distinguished from the Illinoian till.

LATE WISCONSINAN DEPOSITS

The most recent glaciation, of late Wisconsinan age, reached its maximum position between 21 and 20 ka radiocarbon years before present. Because of the short time available for weathering and erosion since retreat, the effects of this glaciation are much more pronounced than those of the earlier two. Late Wisconsinan deposits are unweathered or very lightly weathered. Clasts of carbonate rock are decayed only where they are not abundant, and some gneiss clasts may have a thin weathering rind, but other clasts are unweathered. The deposits retain original depositional morphology, except on very steep slopes where they are eroded or where they have been trimmed or gullied by streams (fig. II-7).

Glacial Advance

Striations, drumlin axes, and the provenance of late Wisconsinan till sheets, indicate that late Wisconsinan ice advanced across northern New Jersey in two main lobes (fig. II-2): a southerly to southeasterly flowing lobe that advanced across the Valley and Ridge and Highlands Provinces and deposited the segment of the terminal moraine west of Denville; and a southerly to southwesterly flowing lobe in the Newark Basin. The Watchung

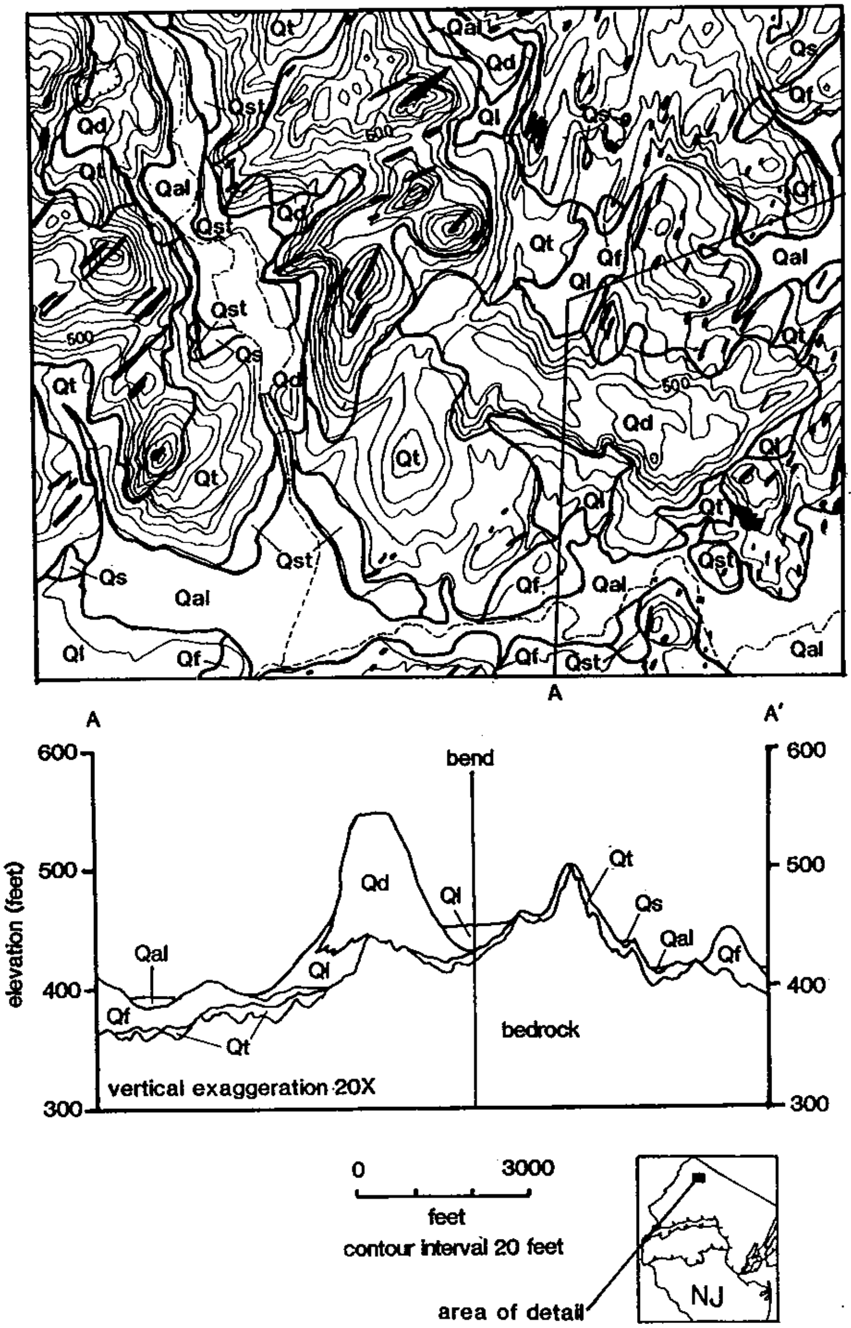


Figure II-7. Map and section of typical surficial deposits within the area of late Wisconsinan glaciation near Sussex, New Jersey. Units are: Qt=till, Qd=deltaic sand and gravel, Ql=lake-bottom silt and clay, Qf=lacustrine-fan sand and gravel, Qs=swamp deposits, Qal=alluvium, Qst=postglacial stream-terrace deposits. Unlabelled areas are thin, discontinuous till. Black areas are bedrock outcrops. Topography from U. S. G. S. Hamburg quandrangle. Geology modified from Stanford and others (1998). Note the largely uneroded glacial landforms, abundant bedrock outcrop, and absence of colluvium.

Mountains divide this lobe into two sublobes: one in the Passaic lowland west of the Watchungs, which deposited the segment of the terminal moraine between Denville and Summit, and a larger sublobe centered in the Hackensack lowland between the Watchungs and the Palisades Ridge, which deposited the large looping moraine east of Summit. The southernmost point of this moraine segment, at 40°30' at Perth Amboy, is the southernmost glaciated point east of the Ohio Valley. The absence of cross-cutting moraines, overlapping till sheets, and crossing striations along the lobe boundaries indicate that they advanced nearly simultaneously. This conclusion, based on detailed field mapping, contrasts with previous hypotheses that advocated different ages for different segments of the terminal moraine, based on variation in the degree of weathering of carbonate clasts (MacClintock, 1954), and on interpretation of the interlobate reentrant at Denville as a younger moraine to the northeast overprinting an older moraine to the west (Connally and Sirkin, 1973). The maximum extent of the late Wisconsinan glacier is marked by the terminal moraine, which is a continuous belt of till between 50 and 200 feet thick forming ridge-and-swale topography. In a few places, the limit is marked by nonmorainic late Wisconsinan till that extends as much as one mile beyond the southern edge of morainic topography.

During advance, nonresistant rocks, including carbonate rocks in the Valley and Ridge and Highlands and certain sandstone and shale units in the Newark Basin, were locally scoured to depths of as much as 350 feet below preglacial valley bottoms. Resistant rocks, chiefly quartzite, gneissic granite, basalt, and diabase, were abraded on slopes facing advancing ice and quarried on lee slopes, although little overall ridgetop lowering is evident, given the similar summit elevations within and outside the glacial limit. Outcrop is extensive on these uplands, in many places comprising 50% of the surface area in the Valley and Ridge and Highlands. Carbonate rocks in the Valley and Ridge also have extensive outcrop where they rise above valley-fill deposits. Locally, however, ice did not erode. In places, preglacially weathered gneiss and carbonate rock as much as 200 feet thick are preserved beneath till, and weathered marble extends to a depth of 1900 feet at the Sterling Hill zinc mine at Ogdensburg.

Till was deposited in hillside ramps, drumlins, and sheets. The ramps are on slopes facing advancing ice, where till may be as much as 150 feet thick. Till in drumlins may be as much as 200 feet thick, although some drumlins have cores of weathered gneiss, Illinoian till, or stratified drift. Till sheets, which are most extensive in the Newark Basin, are generally 20 to 40 feet thick. In some valleys just back from and beneath the terminal moraine, till is stacked atop Illinoian deposits and proglacial late Wisconsinan lacustrine deposits, indicating overramping rather than erosion in this marginal zone.

Radiocarbon Dates

Relevant ^{14}C dates bracketing the arrival and retreat of the late Wisconsinan glacier are listed in table II-1 and shown on a time-distance plot in figure II-8. Radiocarbon ages from the core taken at Budd Lake provide a possible maximum date for the arrival of late Wisconsinan ice at the terminal moraine. Harmon (1968) obtained a date of 22,870±720 yrs BP (I-2845) from clay at 37 feet in the core, within an interval of dominantly pine (50-60%) and spruce (10-20%) with some oak (5-10%) and Ambrosieae dominant in the non-arboreal pollen. This date is below the apparent peak cold at 30-32 feet, where spruce peaks at about 70%, oak is absent, and sedge and grass show peaks in the non-arboreal count. A second date of 12,290±570 yrs BP (GXO-330) at 27 feet was obtained on gytta at the top of the spruce zone. Harmon rejected the dates as too old based on correlation of the upper date to zones elsewhere in the northeast, and attributed the excess age to contamination by old carbon from Paleozoic carbonate rock, although the drift around Budd Lake contains only trace amounts (less than 1% in the gravel fraction) of carbonate rock. The valley fill at the north end of Budd Lake, as inferred from water-well records, shows Illinoian deposits to an elevation of about 900 feet, sufficient to maintain a lake in the Budd Lake basin at the elevation of the lower sample (about 890 feet) from the Illinoian deglaciation to the arrival of late Wisconsinan ice. Thus it is possible that the lower date is uncontaminated and precedes arrival of late Wisconsinan ice.

A date of 20,180±500 (QC-1304) (Stone and others, 1989) on a concretion from glaciolacustrine sediments south of the moraine in the Lake Passaic basin date ice at or shortly after retreat from the terminal moraine there. Minimum dates for deglaciation include basal postglacial dates of 19,340±695 (GX-4279)

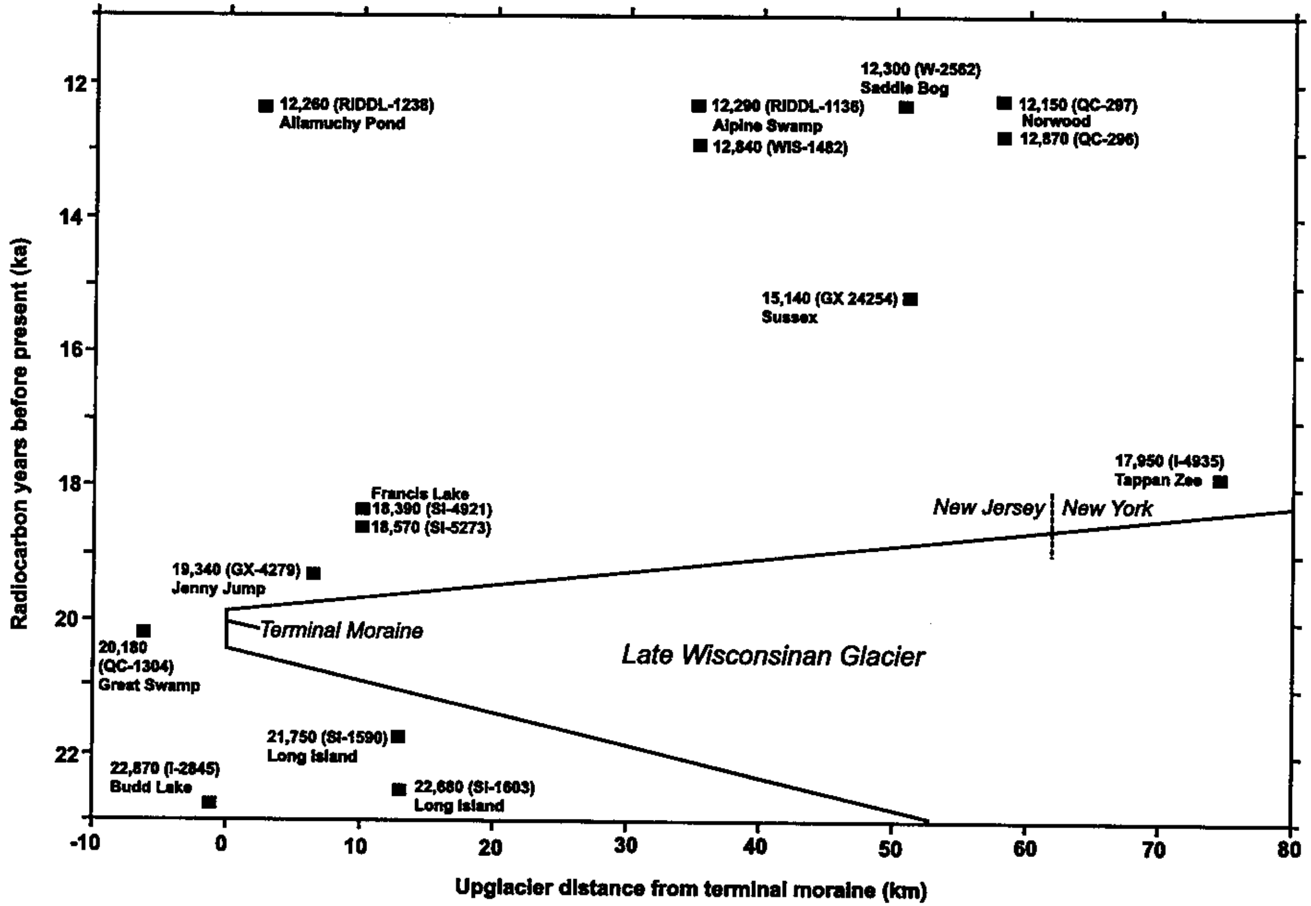


Figure II-8. Time-distance diagram showing radiocarbon dates for the late Wisconsinan glaciation of New Jersey. Refer to Table II-1 for details.

Table II-1. Radiocarbon dates bracketing the Wisconsinan glaciation of New Jersey.

Location	Material	Date	Reference	Stratigraphy
Budd Lake 40°52'30"; 74°45'05" depth=11.3 m	organic lake clay	22,870±720 (I-2845)	Harmon (1968)	In pre-advance deposits, in a spruce pollen horizon 2 m below herb zone
Port Washington, Long Island depth=9 to 27 m	shells, peat, wood, organic silt	21,750±750 (SI-1590) on organic silt 22,680±440 (SI-1603) on peat	Sirkin and Stuckenrath (1980)	Pre-advance dates from deformed beds underlying late Wisconsinan till. 27 additional dates range from 25 to >43.8 ka.
Great Swamp depth=18-28 m	concretions in laminated silt and clay	20,180±500 (QC-1304)	Reimer (1984)	In lower part of varve section in southern basin of Lake Passaic; may date ice at terminal moraine.
Jenny Jump Mt. 40°54'21"; 74°55'08" depth=4.9 m	organic material	19,340±695 (GX-4279)	D. H. Cadwell (personal communication)	Base of postglacial bog. Minimum deglaciation date.
Francis Lake 40°58'15"; 74°50'23" depth=9 m	organic clay	18,570±250 (SI-5273) at 9 m 18,390±200 (SI-4921) at 9 m 16,480±430 (SI-5274) at 8 m	Cotter and others (1984)	Base of postglacial lake deposits, in herb zone. Oldest date was replicated in second core. Minimum deglaciation dates.
Tappan Zee, NY depth=16 m	varved clay and silt	17,950±620 (I-4935)	Weiss (1971)	Near top of glacial-lake varves 33 m thick over till. Minimum deglaciation date.
near Sussex 40°12'32"; 74°33'59" d=1.2 m	clast of organic clay	15,140±470 (GX-24,254)	Stanford (unpub.)	In colluvium along base of delta front. With spruce zone pollen (G. Brenner, written communication). Postdates draining of glacial Lake Wallkill. Minimum deglaciation date.
Norwood 40°59'4"; 73°57'44" d=1 m	peat	12,870±200 (QC-296) at d=1 m 12,150±210 (QC-297) at d=0.8 m	Averill and others (1980)	In wetland sediment overlying a postglacial stream terrace deposited after draining of glacial Lake Hackensack but prior to rebound. With spruce-zone pollen.
Alpine Swamp 40°58'; 73°55' d=10 m	organic clay and plant remains	12,840±110 (WIS-1482) on organic clay at 8.25-8.32 m 12,290±440 (AMS RIDDL-1136) on spruce needle at 9.9 m.	Peteet and others (1990)	In postglacial pond deposit. Upper sample in spruce zone; lower sample in herb zone. Minimum deglaciation date.
Allamuchy Pond 40°55'; 74°50' d=9 m	spruce needle	12,260±220 (AMS RIDDL-1238)	Peteet and others (1993)	In postglacial lake deposit. With spruce-zone pollen. Minimum deglaciation date.
Saddle Bog (on Kittatinny Mt.) 41°14'08"; 74°24'10" d=5.5 m	peat	12,300±300 (W-2562)	Sirkin and Minard (1972)	In postglacial bog deposits. With spruce-zone pollen. Minimum deglaciation date.

(Cadwell, personal communication) from a bog on Jenny Jump Mountain just north of the terminal moraine in the Kittatinny Valley, and 18,570±250 (SI-5273) (Cotter and others, 1986) from Francis Lake, about 10 miles north of the moraine in the Kittatinny Valley; and a date of 17,950±620 (I-4935) (Weiss, 1971) from glacial Lake Hudson sediment in the Tappan Zee area just north of the New York state line. This date, in combination with recessional ice-margin positions (fig. II-10), indicate the late Wisconsinan ice margin had retreated north of New Jersey by about 18 ka.

Erosional Features

On uplands north of the terminal moraine, landscape detail has been markedly shaped by glacial erosion. Freezing of material onto the glacier base removed much of the soil, weathered rock, colluvium, and earlier glacial deposits to expose the underlying bedrock. Once exposed, the bedrock itself was erosionally shaped. On hillslopes that faced advancing ice, resistant rock was abraded as debris-rich ice was forced against the slope, causing pressure-melting and sliding of the basal ice. On hillslopes facing away from the advance direction, rock was quarried when the subglacial water produced by pressure melting flowed into fractures in the rock and then refroze onto the glacier bed. Glacier flow then entrained and removed the ice-encased blocks of rock. Quarrying produced cliffs and steep, blocky outcrops. These ledge-cliff features are common landforms today over large areas of the Highlands and Valley and Ridge provinces, and locally in the Palisades and Watchung Mountains.

Late Wisconsinan Till

In New Jersey most till is locally derived. The grain size and color of the till reflect that of the bedrock immediately upglacier of the site of deposition (fig. II-9a, b). The late Wisconsinan till is divided into three lithic units. Reddish brown sandy-silty till, with few boulders, derived chiefly from red shale and sandstone in the Newark Basin (formally designated as the Rahway Till in Stone and others [in press], number 1 on fig. II-9b) covers most of the Newark Basin. It includes deformed blocks and lenses of Cretaceous clay and sand where it covers the feather edge of the Coastal Plain near Perth Amboy, and includes a silty, yellow to reddish yellow to gray phase where derived from basalt and diabase bedrock (number 2 on fig. II-9b). Light gray to yellowish brown sandy till, with many boulders, derived chiefly from gneiss (the Netcong Till, number 3 on fig. II-9b) covers most of the Highlands and part of the northwestern Newark Basin.. Olive brown to grayish brown sandy-silty till, with few to some boulders, derived from Paleozoic sedimentary rock (the Kittatinny Till, number 4 on fig. II-9b) covers the Valley and Ridge and small parts of the westernmost Highlands. On Kittatinny Mountain, the Kittatinny Till is a yellow to reddish brown silty sand, with many boulders, derived from quartzite and red sandstone (number 5 on fig. II-9b).

Recessional Deposits

As ice stood at the terminal moraine and melted back from it, sand, gravel, silt, and clay (collectively termed "stratified drift") were deposited by meltwater in glacial river plains and glacial lakes. Stratified drift was also deposited during glacial advance, but, except in a few valleys along the terminal moraine, most of these deposits were later eroded by the glacier. In places, till and nonstratified, bouldery debris-flow sediment were deposited in moraines along recessional ice margins, marking the position of the glacier front as it retreated northward. Figure II-10 shows the recessional ice margins, major glacial lakes, and major glacial rivers during the retreat of the late Wisconsinan glacier.

Recessional Ice Margins

Recessional ice margins (fig. II-10) are marked by four till moraines and by the ice-contact heads of fluvial plains and glacial-lake deltas. The moraines are well developed on Kittatinny Mountain (Minard, 1961; Witte, 1997) and are discontinuous but readily traceable in the Kittatinny and Minisink Valleys and the western Highlands (Herpers, 1961; Stanford, 1993; Witte, 1997), but are only locally developed east of there. The late Wisconsinan stratified drift is named the Rockaway Formation in Stone and others (in press). These fluvial and lacustrine deposits, which are divided into about 100 informal allostratigraphic units in Stone and others (in press) that express the lake basin (or group of basins) and fluvial system in which they were laid down, fill most valleys and are as much as 250 feet thick.

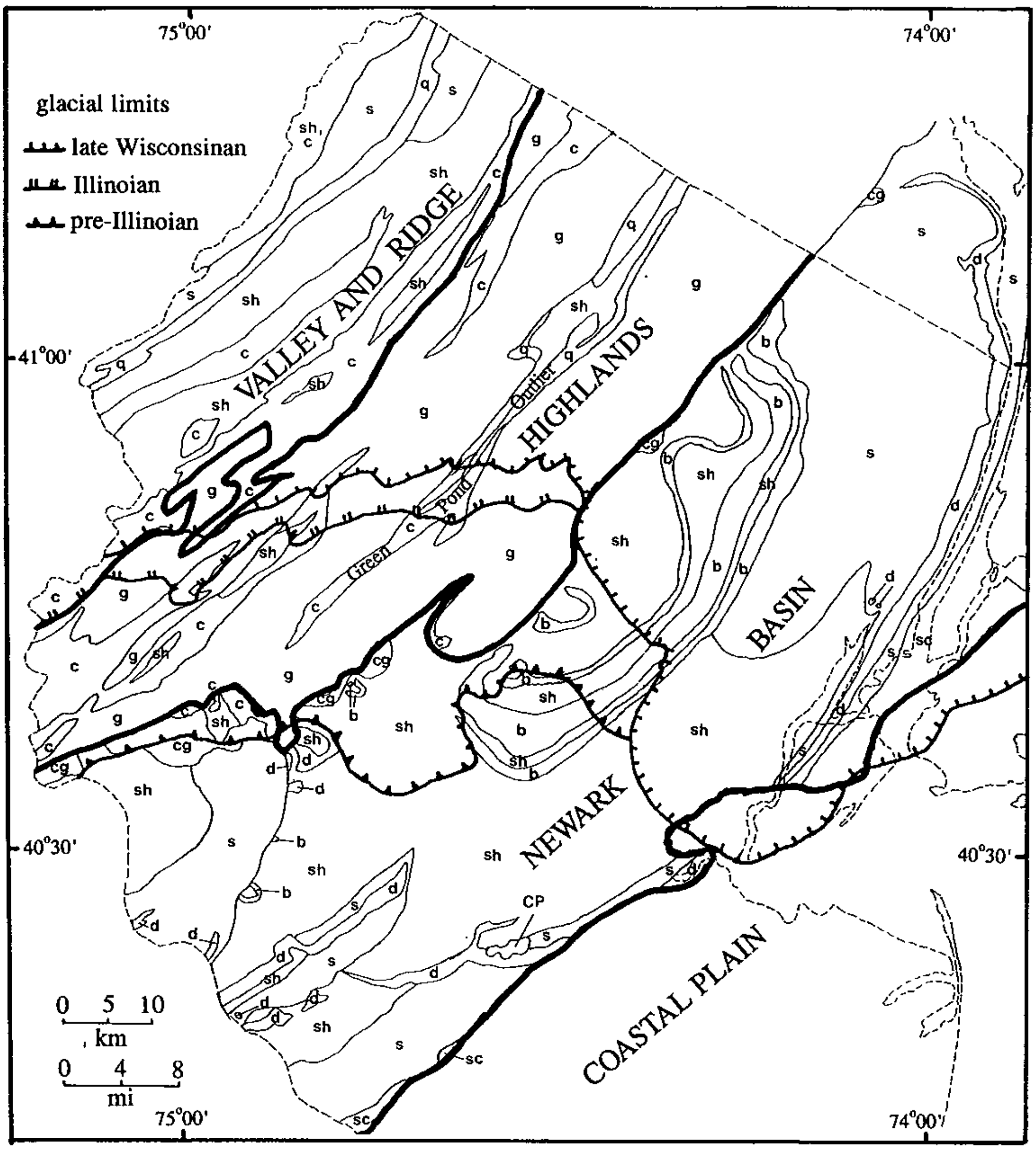


Figure II-9a. Bedrock lithologies of northern New Jersey. Abbreviations are: b=basalt, c=carbonate rock, cg=conglomerate, CP=Coastal Plain sediments, d=diabase, g=gneiss, q=quartzite, s=sandstone and mudstone, sh=shale, sc=schist. Valley and Ridge rocks are Cambrian through Devonian in age; the Highlands include Proterozoic gneiss and marble and Cambrian through Devonian sedimentary rocks; Newark Basin rocks are of Triassic and Jurassic age. Coastal Plain sediments are Cretaceous through Miocene in age.

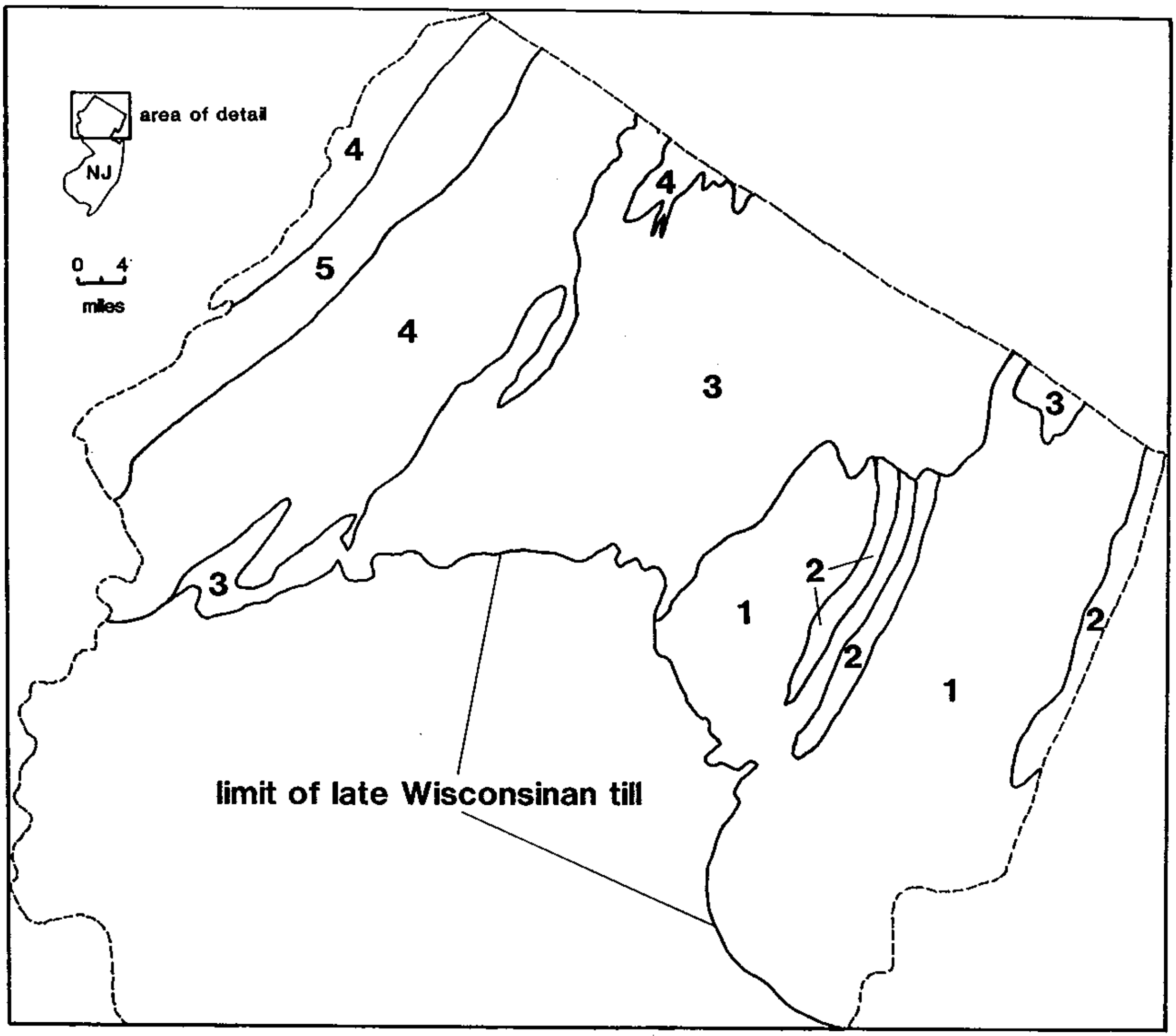


Figure II-9b. Late Wisconsinan till of New Jersey. 1=reddish-brown silty till derived from red shale and sandstone (Rahway Till), 2=yellow to reddish yellow silty till derived from basalt and diabase (Rahway Till, yellow phase), 3=yellowish brown to gray sandy till derived from gneiss (Netcong Till), 4=olive brown to gray silty till derived from carbonate rock and gray slate and sandstone (Kittatinny Till), 5=yellow to reddish brown silty sand till derived from quartzite and red sandstone and shale (variety of Kittatinny Till).

Packages of deltaic and fluvial sediment deposited from specific ice-margin positions are known as "morphosequences" (Koteff and Pessl, 1984). Fluvial morphosequences are identified by mapping terraces that grade to, and texturally coarsen toward, ice-contact slopes marking the ice margin from which the plain was deposited. Fluvial morphosequences typically shingle upvalley, marking successive recessional margins. Deltaic morphosequences are identified by matching the elevation of the delta-top, or, ideally, the contact of the topset and foreset beds within the delta, to the elevation of the spillway for the lake basin. Identifying the spillway and corresponding delta constrains the position of the ice margin required to dam the lake. In smaller lake basins, deltaic morphosequences typically step down to the north as the retreating ice front uncovers successively lower outlets, or as spillways erode. Delta elevations may rise slightly to the north if sediment dams across the valley create successively higher spillways, and in large lakes where postglacial rebound has elevated northerly portions of the former lake basin relative to southerly portions. Mapping morphosequences thus provides a way to reconstruct recessional ice margins when moraines or clear ice-contact features are absent.

The history of lake levels and drainage, determined from morphosequence mapping and from the locations of spillways and meltwater channels, indicates stepwise northward retreat of a single ice margin and, in combination with the moraines, constrains the position of that margin. Retreat occurred without significant readvance. Minor readvances of less than 1 or 2 miles at several locations in the Passaic and Hackensack lowlands are marked by till that overlies and deforms lake sediments. These occur near Little Falls and Preakness in the northeastern part of the Lake Passaic (Salisbury, 1902) and in Lake Bayonne near Elizabeth (as inferred from logs of borings). Details of the deglaciation are provided by Ridge (1983), Witte (1988, 1997), Stone and others (1989), Stanford and Harper (1991), Stanford (1993b), and Stone and others (in press), and on quadrangle maps available from the New Jersey Geological Survey.

Glacial Lakes and River Plains

Valleys that sloped away from the glacier margin and were not dammed by previously-deposited glacial sediment conducted streams of glacial meltwater, which deposited glaciofluvial plains. These plains were deposited as ice stood at the terminal moraine in the Delaware and Musconetcong valleys and in valleys near Plainfield, Metuchen, and Perth Amboy (indicated on figure II-10 by the "glacial stream" symbol). During recession, fluvial plains were deposited in the Wanaque, Pequannock, Rahway, Elizabeth, and Saddle river valleys, and in several small tributary valleys (fig. II-10). The total thickness of the deposits forming these plains is generally less than 50 feet.

Glacial lakes occupied many more valleys than glacial streams did, and most stratified drift is glaciolacustrine. Glacial lakes in New Jersey formed where (1) valleys sloping away from the glacier margin were dammed by previously-deposited sediment, or where (2) valleys sloping toward the margin were dammed by the glacier. Lakes formed in the first setting are known as sediment-dammed lakes; those formed in the second are known as bedrock-dammed lakes.

Large sediment-dammed lakes formed where major valleys and lowlands were blocked by dams of glacial sediment. They generally have a single spillway, and are characterized by sand and gravel deposited in deltas and lacustrine fans alternating with low-lying lake-bottom plains of silt and clay. This group (fig. II-10) includes Lakes Pequest, Hopatcong (which persists today), Passaic, Bayonne, and Hudson, all dammed by the terminal moraine; and Lakes Beaver Run, Swartswood (which persists today), Big Springs, Tenakill, and Paramus, all dammed by previously-deposited deltas or till.

Small sediment-dammed lakes formed in narrow valleys. Here, successive dams were deposited at sequential recessional ice margins. Each recessional deposit formed a dam for a small lake basin as the ice margin receded northward. Because the lakes were small and their spillways unstable, they quickly filled with sediment. Where they filled completely, glacial streams flowed across the filled-lake surface, depositing a small outwash plain. Series of such lakes formed in the Delaware, Flatbrook, Paulins Kill, upper Pequest, Rockaway, and Ramapo valleys (fig. II-10).

Large bedrock-dammed lakes formed in major valleys and lowlands that sloped towards the glacier margin. They generally have single spillways and, like the large sediment-dammed lakes, are marked by deltas and lacustrine fans alternating with lake-bottom plains. Examples (fig. II-10) include Lakes

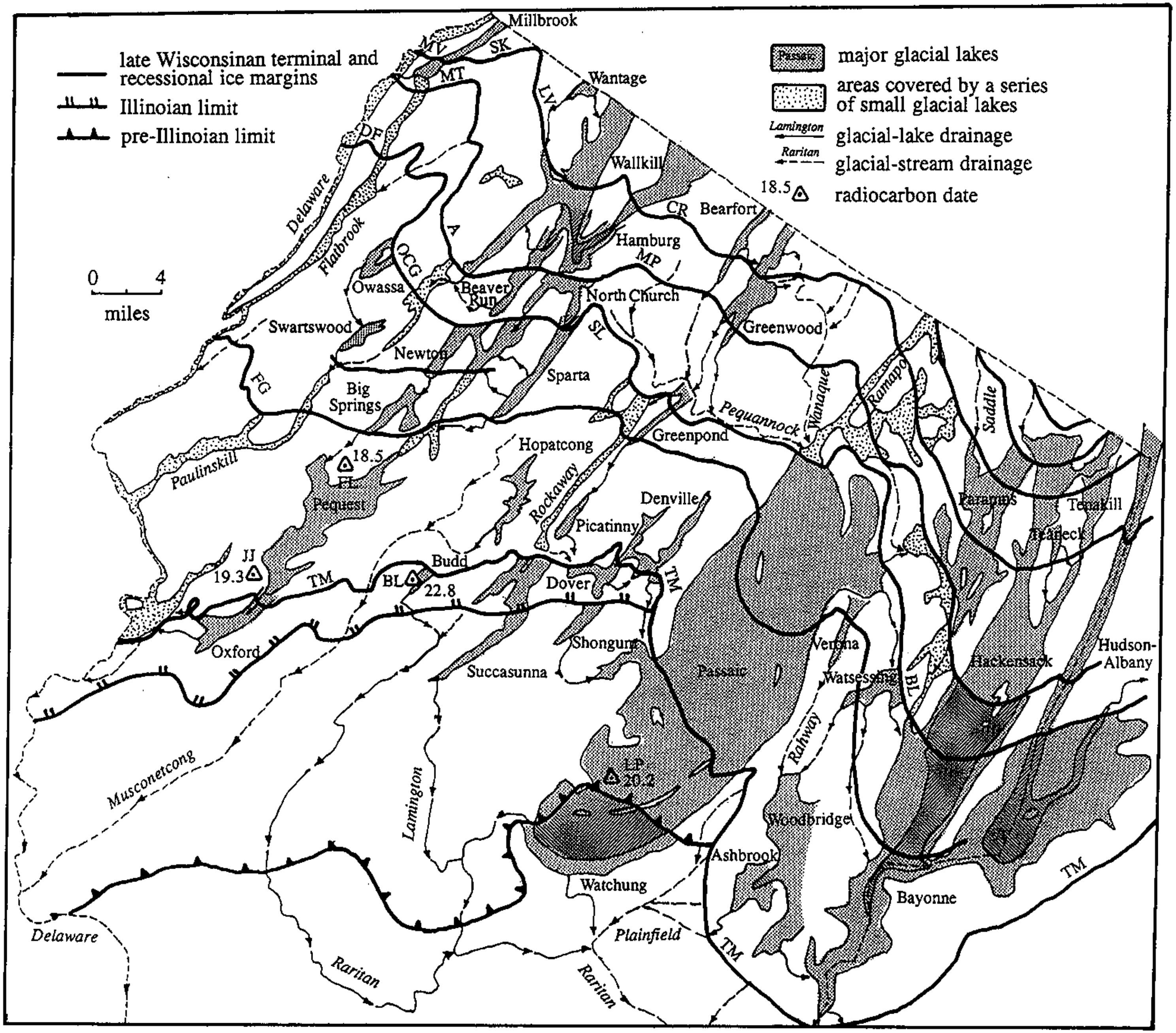


Figure II-10. Glacial limits, late Wisconsinan recessional ice margins, selected radiocarbon date locations, glacial lakes and meltwater drainage. Numerous minor recessional ice margins and small lakes not shown. Lakes Oxford, Budd, Succasunna, and Passaic likely existed during both the Illinoian and late Wisconsinan glaciations. Lake Watchung is a possible pre-Illinoian lake. Abbreviations on ice-margin lines indicate till moraines: TM=terminal moraine, FG=Franklin Grove, DF=Dingmans Ferry, OCG=Ogdensburg-Culvers Gap, SL=Silver Lake, BL=Bloomfield, MT=Montague, A=Augusta, MP=Mud Pond, MV=Millville, SK=Steeny Kill Lake, LV=Libertyville, CR=Cherry Ridge. Radiocarbon date locations are: BL=Budd Lake, JJ=Jenny Jump, FL=Francis Lake, LP=Lake Passaic. Compiled from Ridge (1983), Witte (1988, 1997), Stone and others (1989), Stanford and Harper (1991), Stanford (1993b), and Stone and others (in press).

Hackensack, Wallkill, Succasunna, Newton, Sparta, Millbrook, Hamburg, Owassa, Oxford, Budd, Greenwood, Verona, Ashbrook, Watsessing, Woodbridge, Bearfort, Greenpond, and Teaneck. Lakes Owassa, Budd, Greenwood, and Greenpond persist in reduced form today.

Small bedrock-dammed lakes formed in tributary valleys that sloped toward the retreating ice margin. They are generally located on uplands between the larger lakes. Here, a series of impoundments formed, controlled by successively lower spillways uncovered as the ice margin retreated. Deposits in these lakes generally include only small deltas, which in places fill the entire lake basin. Lake-bottom plains generally are absent.

Thickness of the stratified deposits in glacial lakes is variable. Deltaic and lacustrine-fan sand and gravel may be as much as 200 feet thick but commonly are 50 to 100 feet thick. Lake-bottom silt, clay, and fine sand are as much as 250 feet thick in places in the Lake Passaic and Lake Hackensack basins (fig. II-4) but commonly are 20 to 150 feet thick.

Drainage Dislocations

Deposition of the valley-fill deposits, and of till, caused a number of alterations to preglacial drainage patterns. Among the most prominent are the lower Raritan, which was diverted southeastward into Raritan Bay when its northeast-trending preglacial valley from Bound Brook to the Elizabeth area was filled by the terminal moraine and Plainfield glaciofluvial plain (Stanford, 1993a). The diverted river cut a narrow gorge through shale at New Brunswick, in contrast to the much broader valley upstream from the point of diversion. The upper Passaic basin, including the Rockaway and Whippany valleys, formerly drained to the preglacial Raritan through the gaps in First and Second Watchung Mountains at Millburn. The gap in Second Mountain was filled with till of the terminal moraine and deltaic sediment of Lake Passaic, forcing the postglacial Passaic to flow north, join the Pequannock, Wanaque, and Ramapo rivers, and exit the Watchungs at Little Falls and Paterson. The focusing of these four rivers on the flat, clayey former lake bottom is a major reason for the extensive flooding experienced upstream of Little Falls.

Similarly, but on a smaller scale, the Musconetcong drainage basin at and above Lake Hopatcong formerly drained to the Rockaway via a filled valley that leads from the south end of Lake Hopatcong to the Ledgewood-Kenvil area (Stanford and others, 1996). This valley was blocked by till of the terminal moraine, and postglacial drainage of the Hopatcong basin is now down the Musconetcong valley. In the upper Lamington valley the present southward drainage is reversed from the north-draining preglacial valley, which was filled with lacustrine and moraine deposits. Similar reversals in north-draining valleys filled with lacustrine deposits also occurred at Budd Lake, Green Pond, and possibly in the upper part of the Paulins Kill basin.

Lesser dislocations, where segments of rivers are shifted from their preglacial routes but remain in their main valleys, occurred on the Rockaway at Dover and Denville, the Wanaque at Greenwood Lake, and the Pequannock at Charlotteburg. At these sites, till filled preglacial valley segments, shifting the postglacial rivers into former tributary valleys or across bedrock spurs. Similar dislocations, with stratified drift acting as the valley-filling material, occur on the Delaware at Belvidere, on the Ramapo at Pompton Lakes, on the Wanaque at Wanaque, on the Second River at Bloomfield, and on the Wallkill between Franklin and Hamburg. On much of the shale and carbonate bedrock in the Valley and Ridge and Newark Basin, glacial scour has largely destroyed preglacial valley topography north of the terminal moraine, so pre-late Wisconsinan drainage patterns cannot be reconstructed.

Resources

The late Wisconsinan stratified deposits, because they are thick and fill valley bottoms, are among the most productive aquifers in the state (refer to chapter IV). The sand and gravel have been extensively mined, in many places to the point of entirely removing some deposits. A number of pits are still active, chiefly in western Morris, Sussex, and Warren counties. Most pits east of this area have ceased operations and are now urbanized. Lake clays were formerly mined for brickmaking, particularly in the Lake Hackensack basin at Little Ferry and in the Lake Passaic basin at several locales. All of the clay pits had closed and been urbanized by the 1960s. Peat has also been mined at several locations for use as a soil conditioner.

POSTGLACIAL EVENTS

Compared to the work done by glaciers, the landscape has been only slightly modified since the late Wisconsinan ice margin retreated north of New Jersey about 18 ka. As the ice front receded from valleys, the glacial lakes, unless maintained by sediment dams, drained. Glaciofluvial plains no longer carried meltwater. Streams adjusted to declining discharge and sediment loads and to changes in valley gradient due to rebound. Fluvial and shallow-water delta sand terraces were deposited on parts of the drained lake-bottoms of glacial lakes Wallkill, Pequest, Passaic, Paramus, and Hackensack.

In the northern Lake Hackensack basin there are two postglacial stream terraces. The terrace gradients indicate that, before rebound, drainage in the Hackensack lowland, including the lower Passaic and Saddle rivers, was northeasterly to the Hudson via Sparkill Gap, a deep notch in the Palisades just north of the state line. Rebound reversed the valley gradient, forming a south-sloping terrace along the present route of the lower Passaic and Hackensack rivers (Stanford and Harper, 1991). A radiocarbon date of 12,870±200 yrs BP (QC-297, Averill and others, 1980) on a basal peat on top of the pre-rebound terrace suggests that the onset of rebound was delayed until about that time

Similar northeasterly drainage occurred from the Delaware Valley at Trenton down the lower Millstone Valley, as marked by the continuity of the late Wisconsinan glaciofluvial deposit of the Delaware Valley across the low divide at Port Mercer and down the Millstone. A date of 10,430±160 (I-16,554, Stanford, 1993a) on basal peat on top of the glaciofluvial deposit on the Delaware-Millstone divide indicates rebound-induced swamping and abandonment of this drainage at or somewhat before that date.

After entrenchment into glacial terraces, to depths of as much as 100 feet in the upper Delaware Valley but generally less than 50 feet elsewhere, streams established floodplains. The postglacial floodplains are generally narrow but north- and northeast-draining streams such as the Millstone, upper Passaic, lower Raritan, and Wallkill have wide floodplains, in part because rebound has reduced their gradients, allowing lateral channel migration. In ponds, marshes, and swamps, which generally occupy former glacial-lake bottoms or scoured rock basins on uplands, silt and clay, followed by peat, have accumulated.

As sea level rose during glacial melting, estuaries formed along the lower reaches of the Raritan, Passaic, Hackensack, Hudson, and Delaware valleys, and the river valleys of the Coastal Plain. Rising sea level entered the Hudson Valley, which had been deeply scoured by glacial erosion, before 12 ka (Newman and others, 1969). Estuarine silt in the Hudson Valley is as much as 250 feet thick. The rising sea had also entered the Raritan Valley, where estuarine deposits are as much as 100 feet thick, by about 11.5 ka, based on a date of 11,420±560 (GX-21,687, Stanford, unpublished) from basal estuarine deposits at Perth Amboy.

Postglacial hillslope erosion is slight. At the base of cliffs, falling rock has accumulated in aprons of talus. Large talus aprons occur at the base of the Palisades north of Fort Lee, locally at the base of quartzite cliffs on Kittatinny, Green Pond, Copperas, and Kanouse mountains, and at the base of gneiss cliffs on the west face of Wawayanda Mountain. At the base of a few steep hillslopes, till and fragments of bedrock have moved downslope to form small aprons of colluvium. Some of these aprons have ice-contact scarps at their distal margins, indicating that they were deposited just after melting ice uncovered the slope but while ice still occupied the valley bottom.

Fine sand and silt were blown downwind from the surface of large sand plains. In a few places, notably to the southeast of the Plainfield outwash plain and on the east side of the upper Delaware Valley and the glacial Lake Hackensack basin, they accumulated as dunes or as thin blankets of eolian sediment. South of the glacial limit, thin loess blankets are common on the east sides of the Raritan and Millstone valleys, and occur locally in the Delaware Valley. These deposits are also situated downwind from broad glaciofluvial or periglacial river terraces. The distribution of eolian sediment indicates that prevailing winds in glacial and early postglacial time were westerlies (as they are today).

The warming postglacial climate led to vegetation change. Postglacial pollen records show dominantly pine and spruce, with elevated sedge and grass, extending from before the glacial maximum to about 11 ka. This combination of arboreal and nonarboreal pollen suggests a mixed boreal woodland-tundra vegetation. Pine dominates from about 11 to 9.5 ka, with sedge and grass declining, and oak dominates after 9.5 ka (Harmon, 1968; Nicholas, 1968; Sirkin and Minard, 1972; Watts, 1979; Russell, 1980; Cotter and others, 1986; Peteet and others, 1990, 1993; Russell and Stanford, 2000). Peteet and others (1990, 1993) detected a increase in spruce, fir, larch, birch, and alder between 11 and 10 ka in cores from a bog on the Palisades Ridge and from a lake on Allamuchy Mountain that they interpret as indicating a return to boreal climate correlating to the Younger Dryas cool period of northern Europe. Warming was reestablished at sbout 10 ka.

CONCLUSIONS

Three drift sheets, with distinctive weathering characteristics and erosional preservation, are present in northern New Jersey. The oldest, based on its intense weathering, deep erosion, close association to a Pliocene fluvial deposit, and correlation to till in Pennsylvania and the midcontinent, is either late Pliocene (pre-Illinoian K) or early Pleistocene (pre-Illinoian F or G) in age. Moderate weathering, shallow erosion, and correlation to till in New England and to the marine oxygen-isotope record, suggest an Illinoian age for the intermediate drift. The late Wisconsinan drift is dated by radiocarbon; these dates indicate that late Wisconsinan ice reached its southernmost position no earlier than about 22 ka and retreated north of New Jersey by about 18 ka. Fluvial and lacustrine morphosequences and recessional moraines indicate that this retreat was stepwise, without significant readvance or widespread stagnation.

The principal changes to the New Jersey landscape caused by these glaciations include depositional and erosional features. Depositional features include glaciolacustrine and glaciofluvial valley fills as much as 250 feet thick, forming fluvial plains and terraces, deltas, lacustrine fans, and lake-bottom plains, and till as much as 200 feet thick forming moraines, drumlins, and till sheets. Erosional features include extensive tracts of abraded and plucked bedrock outcrop on upland areas of resistant bedrock like diabase, basalt, gneiss, and quartzite; and scoured troughs as much as 350 feet deep in valley bottoms on nonresistant rocks like carbonates, shale, and weathered sandstone.

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CHAPTER III

Late Wisconsinan end moraines, outwash heads, and ice retreat in Kittatinny Valley and nearby uplands, Sussex and Warren Counties, New Jersey: *A view along the margin of the Laurentide ice sheet*

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ABSTRACT

End moraines in western New Jersey are prominent, segmented, arcuate belts of hummocky till that cross Kittatinny Valley and nearby uplands. They include the Terminal Moraine and several recessional moraines deposited 21,000 to 18,000 years ago during the late Wisconsinan substage of the Wisconsinan glacial stage. They all consist of a complex assemblage of small-scale landforms that collectively define areas of ridge-and-trough and knob-and-kettle topography. Their lobate course, till composition, and preferred development of ridge-and-trough topography along their outer edges show they were initially constructed at the margin of an active glacier. The Terminal Moraine and some larger end moraines were also laid down following a readvance, further evidence of their association with active ice. Although end moraines were initially constructed at active glacier margins, their final form is largely a product of stagnation. Apparently, the glacier's terminus became buried by its own debris, which resulted in the formation of a narrow zone of dead marginal ice. Except for moraine-parallel ridges, which may be either push ridges or colluvial ramparts, morainal topographic elements were largely formed through topographic inversion after a complex history of collapse, due to melting of buried ice and resedimentation of supraglacial debris by mass wastage.

Outwash heads of ice-contact deltas also mark the former glacier margin. During retreat of the Kittatinny Valley lobe, proglacial lakes formed in the Paulins Kill, Pequest, and Wallkill River valleys where drainage became blocked by meltwater sediment, moraine, and ice. The histories of these glacial lakes, and ice-retreat positions marked by outwash heads of ice-contact deltas and end moraines, showed that the Kittatinny Valley lobe retreated systematically to the northeast chiefly by a process of stagnation-zone retreat. In addition, redavances are recorded by the Terminal Moraine, Ogdensburg-Culvers Gap moraine, and Augusta moraine. Five ice margins, the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex, have been identified, and they delineate major ice-retreat positions of the Kittatinny Valley lobe. The Culvers Gap, Augusta, and Sussex margins are traceable across the New Jersey Highlands and into the Newark Basin, and all margins except the Sparta margin are traceable westward across Kittatinny Mountain into Minisink Valley. The strong evidence of systematic deglaciation, and the presence of at least three readvances, show that regional or valley-ice lobe stagnation was not a valid style of deglaciation for Kittatinny Valley.

INTRODUCTION

Location and Setting

Kittatinny Valley is a broad lowland in northwestern New Jersey (Fig. 1) that lies in a glaciated part of the Great Valley Section of the Appalachian Valley and Ridge Province. It is underlain by folded and thrust-faulted belts of dolomite, limestone, slate, and sandstone of Lower Paleozoic age (Fig. 2). The valley is further cut by the Pequest River and Paulins Kill, which flow southwest toward the Delaware River, and the Wallkill River, which drains most of the upper part of Kittatinny Valley and flows northeast toward the Hudson River in New York. These rivers chiefly flow along strike-controlled belts of carbonate rock that are mostly dolomite of the Kittatinny Supergroup (Drake and others, 1996). Relief rarely exceeds 300 feet (91 m), rock outcrops are very abundant, and knobby topography is commonplace. Where the underlying rock contains more chert, narrow, strike-parallel ridges have formed. In other places fluvial erosion, dissolution, and glacial erosion have greatly lowered the valley floor by as much as 200 feet (61 m). Most of these areas are underlain by thick deposits of glaciofluvial and glaciolacustrine sediments laid down during the last ice age. Slate, siltstone, and sandstone of the Martinsburg Formation (Drake and others, 1996) underlie interfluves in Kittatinny Valley and the area between Paulins Kill valley and Kittatinny Mountain. Overall, the average elevation here is about 300 feet (91 m) higher than the carbonate-floored valleys, and relief may be as much as 500 feet (152 m). Topography consists of rolling hills of

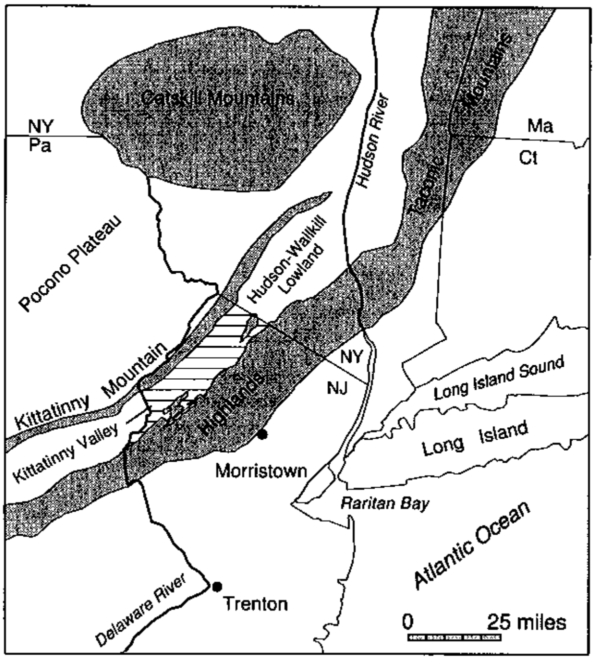


Figure 1. Location of Kittatinny Valley in northwestern New Jersey and regional physiography.

moderate to steep slopes, and many strike-parallel ridges streamlined by glacial erosion. In some places bedrock is buried beneath drumlins and thick ground moraine.

Kittatinny Mountain bounds Kittatinny Valley on its northwest side. The mountain is held up by the Shawangunk Formation, a tough, resistant quartzite, and quartz-pebble conglomerate, and the Bloomsburg Red Beds, which consists of interlayered red sandstone and red shale (Drake and others, 1996). Its nearly level summit forms a continuous ridge that extends from the Shawangunk Mountains in New York southwestward through New Jersey into Pennsylvania. The mountain's main crest rises as much as 1000 feet (305 m) above the valley floor. In places its continuity is broken by large gaps, such as Culvers Gap, and Delaware Water Gap, and several smaller gaps and sags. Topography is rugged and consists of narrow- to broad-crested ridges that trend southwestward paralleling the main trend of the mountain. The mountain's steep southeast face also forms a nearly continuous escarpment in New Jersey. Rock outcrops are very abundant and many have been shaped by glacial erosion. The piedmont that lies to the northwest of the mountain's main ridge is also included as part of Kittatinny Mountain. Bedrock exposures are sparse there because the rock surface is covered by thick ground moraine and drumlins. In the Culvers' Gap area, a series of repeating low amplitude folds and an overall decrease in the northwest dip of the outcrop belt, nearly triples the mountain's width.

The New Jersey Highlands, part of the southern extension of the New England Physiographic province, borders the valley on its southeast side. Included with the Highlands is a large outlier in the southern part of the valley that includes Jenny Jump Mountain, Danville Mountain, High Rock Mountain, and Mount Mohepinoke. These uplands have rugged relief; their rough lands underlain by metasedimentary and intrusive rocks of Proterozoic age that rise as much as 1000 feet (305 m) above the floor of Kittatinny Valley. Ridge lines chiefly follow layering in the country rock. Although, discordant trends are common, and in places deep gaps cut across the southwest-trending topographic grain. Glacially scoured outcrops are common.

The strong northeast-trending topographic grain of the of northwestern New Jersey developed slowly during the Tertiary Period. The development of the valley-and-ridge physiography was largely controlled by the structure of the underlying sedimentary strata and their varying resistance to erosion. Because the high-ridge area of Kittatinny Mountain is underlain by the erosion resistant Shawangunk Formation, it holds up the higher areas. During the Pleistocene Epoch, the action of at least three continental glaciers modified the landscape by deeply scouring valleys, wearing down and smoothing bedrock ridges and hills, and by eroding soil and loose rock. Debris captured by the ice sheets was either deposited as till or carried off and deposited by meltwater streams.

Previous Investigations

The surficial geology of Kittatinny Valley was first discussed by Cook (1877, 1878, and 1880) in a series of Annual Reports to the State Geologist. He included detailed observations on the age, distribution, and kinds of glacial drift, and evidence for glacial lakes. Lewis (1884) traced a terminal moraine westward from Delaware Valley to Salamanca, New York and considered it the same age along its length. A voluminous report by Salisbury (1902) detailed the glacial geology of New Jersey region by region. The Terminal Moraine (Fig. 3) and all glacial drift north of it were interpreted to be of Wisconsinan age deposited during a single glaciation. Kames, kame terraces, deltas, recessional moraines, and glacial lakes were also described in Kittatinny Valley, where Salisbury also noted that "in the northwestern part of the state, several halting places of ice can be distinguished by the study of successive aggradation plains in the valleys." Although

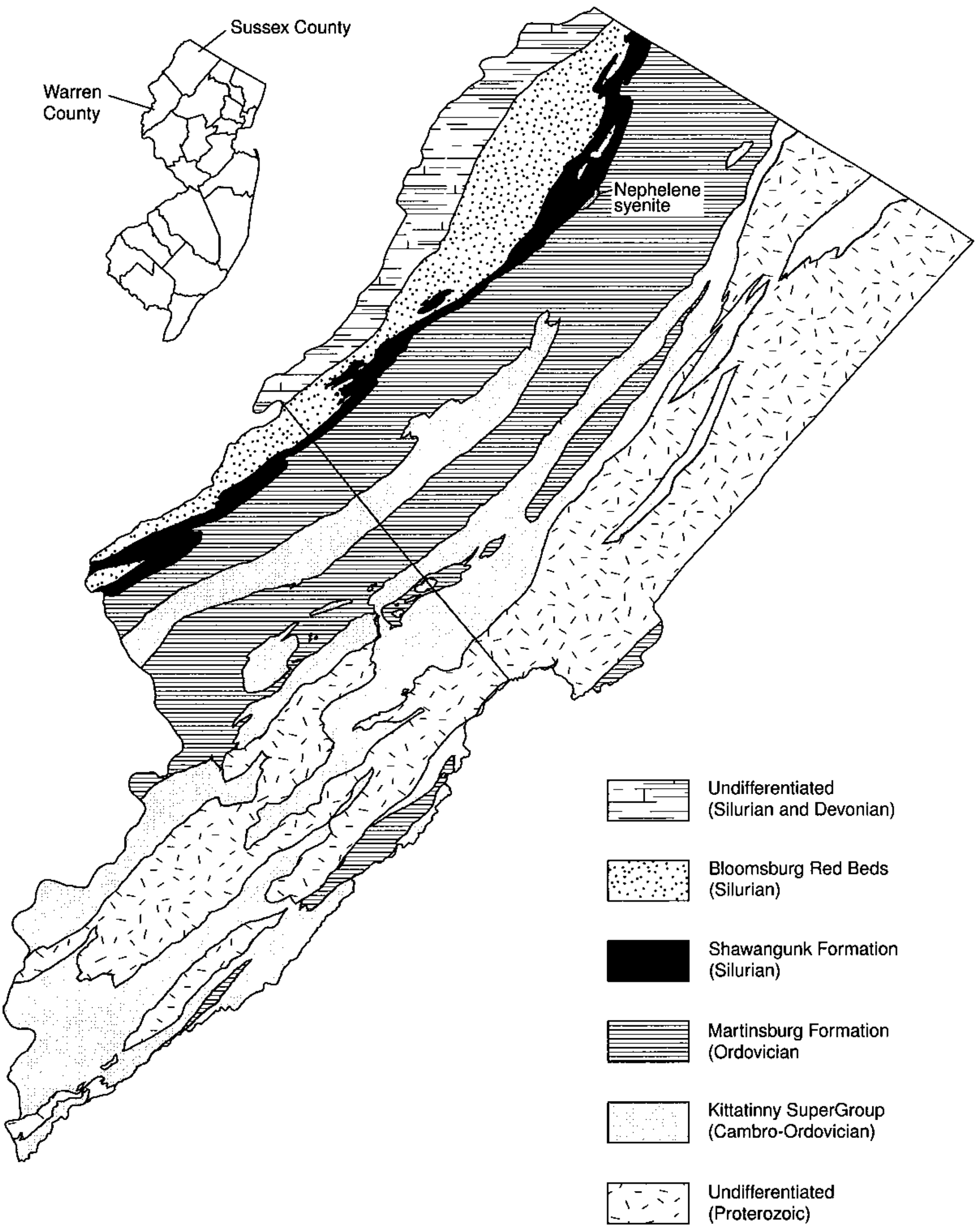


Figure 2. Simplified geologic map of Sussex and Warren Counties, New Jersey. Data modified from Drake and others (1996). Lithologic Key: Proterozoic formations - chiefly metasedimentary and intrusive rocks with minor marble; Kittatinny SuperGroup - dolomite and limestone; Martinsburg Formation - slate, shale, siltstone and graywacke; Shawangunk Formation - quartzite and quartz-pebble conglomerate; Bloomsburg Red Beds - shale and sandstone; Undifferentiated Silurian and Devonian Formations - shale, siltstone, sandstone, and limestone. Due to limited outcrop area the Jacksonburg Limestone is included with the Kittatinny SuperGroup, and the Hardyston Quartzite is included with the Kittatinny SuperGroup.

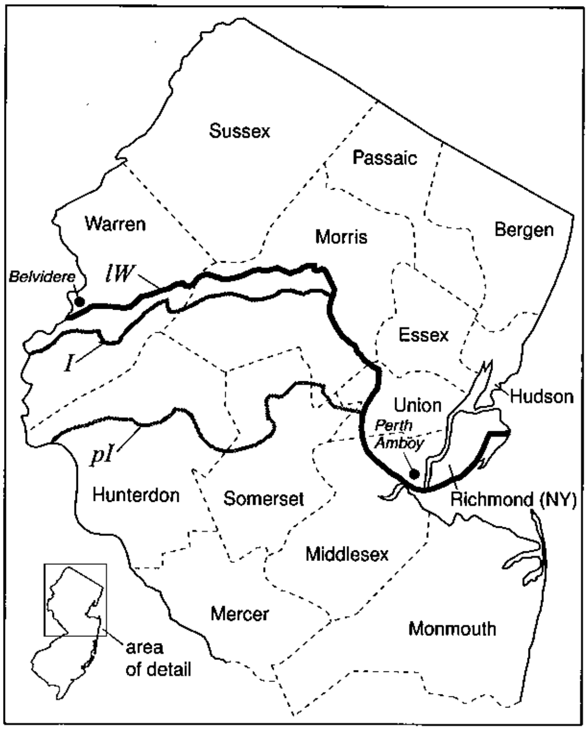


Figure 3. Limits of glaciations in New Jersey and nearby New York. The trace of the lW limit generally marks the position of the Terminal Moraine. Key: *lW* - late Wisconsinan, *I* - Illinoian, and *pI* - pre-Illinoian.

it was recognized that some of these deposits marked ice-retreat positions, they were not documented within a larger chronostratigraphic framework. In Pennsylvania, Leverett (1934) slightly modified the trace of the Terminal Moraine and also assigned a Wisconsinan age to it and the glacial drift north of it. He also suggested that the pre-Wisconsinan drift was laid during the Illinoian and Kansan glaciations. Ward (1938) proposed that an extensive ice tongue occupied the Delaware Valley during the Wisconsinan glaciation, while Kittatinny Mountain near Delaware Water Gap was not covered by ice. MacClintock (1954) divided the glacial deposits into Olean drift of early Wisconsinan age, and the Binghamton drift of late Wisconsinan age, based on the depth of carbonate leaching in glacial stream sediments. Crowl and Sevon (1980) showed that the Wisconsinan deposits consisted only of the Olean drift, which they assigned a late Wisconsinan age and that the older glacial deposits were represented by the Warrensville drift of early Wisconsinan age, and the Muncy drift of Illinoian age. Cotter and others (1986) also showed that the youngest glacial deposits in Pennsylvania and New Jersey are late Wisconsinan age, and are correlative with the Olean drift in Pennsylvania, and Ridge and others (1990) indicated that older and weathered drift in the Delaware Valley north of Marble Mountain is late Illinoian age and not early Wisconsinan. Braun (1989), Witte and Stanford (1995), Stone and others (in press) demonstrated that the youngest glacial deposits in eastern Pennsylvania and New Jersey are late Wisconsinan age, and that the two older drifts are Illinoian, and Pre-Illinoian age. Gardner and others (1994) showed that the Pre-Illinoian drift in central Pennsylvania is older than 788 ka, based on the reversed magnetic polarity of glacial lake-bottom deposits preserved near Antes Fort in the West Branch Susquehanna River valley. Similarly, high standing, weathered, fined grained glaciofluvial deposits south of the Illinoian drift border in Pohatcong Creek valley, western New Jersey, also showed a reversed remnant magnetic polarity (J. C. Ridge, written commun. 1999).

Recessional moraines in Kittatinny Valley were originally identified by Salisbury (1902), and later remapped by Herpers (1961), Ridge (1983), and Witte (1988). Both the Ogdensburg-Culvers Gap and Augusta moraines were traced over Kittatinny Mountain by Herpers (1961), Minard (1961), and Witte (1997a). In Kittatinny Valley, Connally and Sirkin (1973) indicated that the "Culvers Gap" moraine, not the Terminal Moraine, marked the southern limit of the late Wisconsinan ice sheet. Later, Connally and others (1989) suggested that the limit was several kilometers south of the "Culvers Gap" moraine, based on the location of a "dead-ice sink" in the head of Paulins Kill valley. They further added that the moraines at Ogdensburg, Augusta, and Sussex are inversion ridges (meltwater deposits laid down in large cross-valley crevasses in stagnant ice), and therefore, they do not delineate ice-retreat positions. They also proposed that deglaciation occurred primarily by large-scale valley-ice lobe stagnation. Their interpretation was based on their recognition of extensive esker systems, massive crevasse-fill deposits, inversion ridges, and dead-ice sinks in the upper part of Kittatinny Valley. Crowl and Sevon (1980) and Cotter (1983) demonstrated that glacial drift north of and including the Terminal Moraine is all of late Wisconsinan age and see no evidence to support a late Wisconsinan maximum position at or near the Ogdensburg-Culvers Gap moraine. A recent investigation by Larsen and Bierman (1995), whom used cosmogenic ^{26}Al dating of gneiss and quartzite erratics, also indicated the Terminal Moraine is of late Wisconsinan age.

Witte (1988) and Ridge (1983) accepted the late Wisconsinan age for the Terminal Moraine, and demonstrated that deglaciation was systematic in a northeast direction, and chiefly by stagnation-zone retreat. Ridge

(1983) showed that Terminal Moraine was composed of several segments constructed at several ice-margin positions. Witte (1991, 1997a) further added that the "inversion ridges" of Connally and others (1989) at Ogdensburg and Augusta are end moraines, and part of a much larger end-moraine complex that marks major ice-retreat positions of the Kittatinny and Minisink Valley ice lobes.

Preglacial Drainage

The primary drainage routes in the study area were probably established well before the Pleistocene. Culvers Gap and possibly the transverse gaps through Jenny Jump Mountain and the Highlands are relicts of an earlier Raritan River drainage network that may have flowed in a southeasterly direction (Fig. 4). The Delaware River, through headward erosion and stream capture, has enlarged its drainage area chiefly by extending its tributaries upvalley along the strike of less resistant rock. In response to the overall lowering of sea level during the Pleistocene the drainage has further evolved by incision, which along the larger tributaries of the Delaware River has resulted in the formation of a much lower, narrower, river valley. Extensive headward erosion by first and second-order streams has also resulted in the dissection of the older valley floor, and the surrounding Highlands. The low elevation of Illinoian glaciofluvial deposits in the Delaware Valley suggests that the river valleys in the study area had been lowered or nearly lowered to their present levels by the time of the Illinoian glaciation.

END MORAINES

Trace

In western New Jersey, end moraines are distinct, segmented belts of bouldery, hummocky glacial drift (Fig. 5) that consist of poorly compact, sandy, bouldery till, minor lenses of glaciotectonized substrate (outwash, weathered bedrock, older till, and colluvium), and water-laid sand, gravel, and silt. The Terminal Moraine is as much as 130 feet thick and one mile in width. The recessional moraines are as much as 65 feet (20 m) thick, and 2500 feet (762 m) wide. Although, most are less than 1000 feet (305 m) wide. All the end moraines have asymmetrical topographic profiles with their distal slopes (outer edges) being the steepest. Their distal margins also have sharp boundaries, whereas the boundaries of their inner margins are typically indistinct. Morainal topography (sharpness, and overall relief) is also better formed along the moraine's outer margin than its inner margin. This was noted by Salisbury (1902, p. 251) whom observed that "...the characteristic morainic topography made by the close association of hummocks, kettles, ridges, and troughs, is, in general, better marked in the outer half of the moraine belt than in the inner."

The Terminal Moraine follows a nearly continuous looping course through Warren County (Fig. 5). In most places the morainal topography is distinct and easily recognized by its well-formed ridge-and-trough and knob-and-kettle topography. In a few places, where steep topography constrained its formation, morainal topography is only faintly noticeable. R.D. Salisbury and H.B. Kummel (Salisbury, 1902) described in detail the course of the moraine across New Jersey. Their excellent description of its trace through the southern part of Kittatinny Valley and across the Jenny Jump outlier stands today, except for a few areas in the Pequest and Delaware Valleys where outwash was mapped as part of the moraine. Clearly, the moraine's course was strongly influenced by topography, extending more southward in areas of lower elevation, and as it approaches the central axis of Kittatinny Valley. In many places a narrow belt of late Wisconsinan till extends as much as 3000 feet (914 m) out beyond the Terminal Moraine. This shows that the Terminal Moraine in many places does not represent the late Wisconsinan glacial border.

Recessional moraines in Kittatinny Valley include Franklin Grove, Fairview Lake, Ogdensburg-Culvers Gap, Augusta, and Libertyville (Fig. 5). The smaller Fairview Lake and Libertyville moraines are correlated with heads-of-outwash situated farther east. The larger ones are more continuous in the valley, and both the Ogdensburg-Culvers Gap and Augusta moraines are traceable across Kittatinny Mountain where the former joins the Dingmans Ferry moraine and the latter joins the Montague moraine (Witte 1997a).

The Franklin Grove moraine (Fig. 5) was first described by Salisbury (1902) and later named by Ridge (1983). The moraine trends north westward from Spring Valley, through Franklin Grove, toward Sand Pond, and ends abruptly at the base of Kittatinny Mountain. This moraine does not continue across the mountain and it is absent east of Spring Valley. However, it is correlated with the Lake Pequest and Andover Ponds morphosequences situated farther east (Ridge 1983; Witte 1988, 1991) in Kittatinny Valley, and with the Sand Hill Church deposits in Pennsylvania (Witte 1997a).

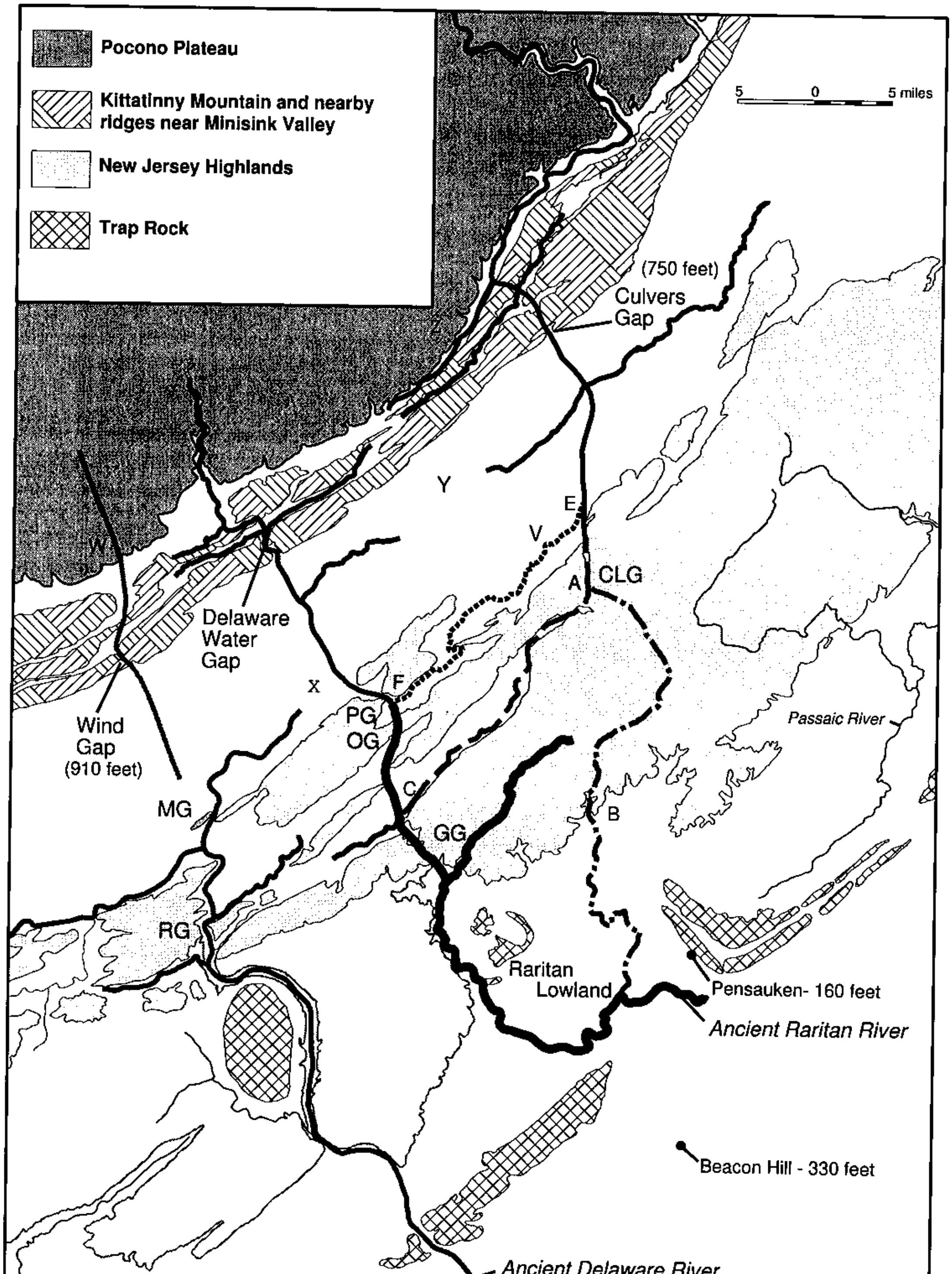


Figure 4. Reconstruction of the late course of the Culvers Gap River, and several scenarios for its capture by the ancient Delaware River. Key to gaps not named on figure: PG - Pequest Gap, OG - Oxford Gap, GG - Glen Gardner Gap, MG - Marble Mountain Gap, RG - Riegelsville Gaps, CLG - Cranberry Lake Gap. Pre-capture course of the Culvers Gap River: A-B Andover-Ledgewood course, A-C Andover-Musconetcong Valley Course, E-F Pequest Valley course. Location of capture: V - Pequest Valley capture, W - Wind Gap capture, X - Pequest Gap capture, Y - Paulins Kill valley capture, Z - Mininsink Valley capture. Elevation of wind gaps and Beacon Hill (Late Miocene) and Pennsauken (Pliocene) Formations are listed. Modified from Witte (1997b).

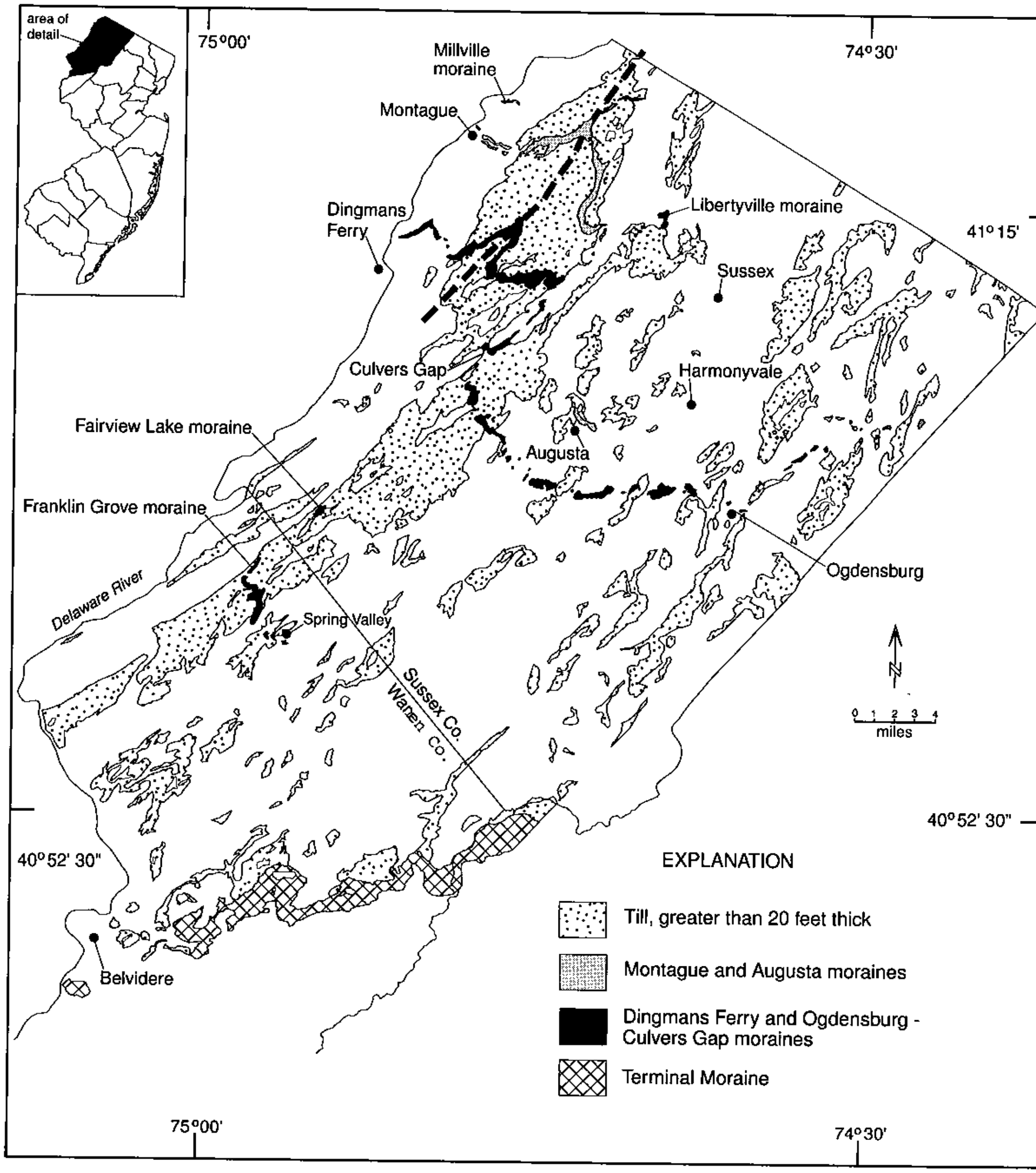


Figure 5. Surficial geologic map showing the late Wisconsinan Terminal Moraine, recessional moraines, and thick till in Sussex and Warren Counties, New Jersey. Dashed line marks the contact between the Dingmans Ferry and Ogdensburg-Culvers Gap moraines, and the Montague and Augusta moraines. Data from Ridge (1983), Stone and others (1989), Witte (1988, 1991), Stanford and others (1990), and Witte and Stanford (1995).

The Ogdensburg-Culvers Gap and Augusta moraines (Fig. 5) were first described by Salisbury (1902) and traced onto Kittatinny Mountain by Herpers (1961), and Minard (1961), and later remapped by Witte (1988, 1997a). The Ogdensburg-Culvers Gap moraine consists of several segments that trend westward from the New Jersey Highlands, through Ogdensburg to Culvers Gap in a distinct cross-valley loop. It continues along the southwest side of Kittatinny Mountain to where it crosses the mountain's main ridge crest, approximately 4 miles (6 km) northeast of Culvers Gap. From here its course traces a smaller loop through the Big Flat Brook valley and joins the Dingmans Ferry moraine. The Augusta moraine consists of several segments that trend westward from the New Jersey Highlands, through Harmonyvale to the base of Kittatinny Mountain, where it lies approximately 3 miles (5 km) northeast of Culvers Gap. From here it can be traced onto Kittatinny Mountain where it joins the Montague moraine, following a course similarly parallel to the Ogdensburg-Culvers Gap moraine. East of Harmonyvale in Kittatinny Valley, morainal deposits are absent or they may lie buried beneath outwash. Ice retreat positions here are marked by ice-contact deltas in the Beaver Run and Wallkill River valleys, and they are correlated with the same ice margin as that marked by the moraine.

Morphology

End moraines consist of a variety of topographic landforms that collectively form a belt as much as one mile wide of thick uneven bouldery drift. The complex assemblage of wetlands, boulder fields, ridges, and mounds provides a diverse ecological setting for many kinds of plants and animals. To the casual observer the collection of ridges, mounds, and depressions appear random and chaotic in their trace across the countryside. Many end moraines have been simply described as a belt of hummocky drift. However, upon close inspection, end moraines consist of several types of topographic elements that can be mapped, and characterized. Thereby providing insight as to their formation, and the nature of sedimentation at the terminus of a continental glacier.

Morainal landforms may be grouped into positive and negative topographic elements (Fig. 6 and 7). Positive elements include ridges, knolls, and plateaus. Ridges are further divided into moraine parallel ridges (MPR) and moraine tangent ridges (MTR). MPR's generally lie along the outer margin of the moraine where they parallel its trace. They have narrow to broad crests, stand as much as 50 feet (15 m) high, and are as much as 2000 feet (610 m) long, although most are less than 500 feet (152 m) long. Many appear to have been formerly continuous, but may have been disconnected by collapse during melting of buried ice. Ridge crests follow straight to slightly arcuate traces that parallel the moraine's outer border. In places they form nested sets that exhibit a remarkable degree of parallelism (Fig. 7), suggesting they were built at several ice-margin positions. Their topographic profiles are typically asymmetric with their inner slopes the steepest. Inner slopes are also hummocky showing that this part of the ridge was laid down against ice. MPR's typically occur along the outer part of the morainal belt. They are either push ridges, formed where the advancing ice had bulldozed ice-marginal sediment, or they are colluvial ramparts, laid down where the glacier margin remained stationary, shedding an apron of debris off its terminus. MTR's are found throughout the morainal belt and they are of similar dimensions as MPR's, although they are not as numerous. Their ridge crests lie tangent to the moraine's course and they follow straight to sinuous traces. Side slopes are typically steep-sided and hummocky. In places the trace of their ridge crests define polygonal patterns. They may be crevasse fillings formed where supraglacial debris had accumulated in deep fractures.

Knolls consist of low, rounded or elliptical hills that vary from larger isolated hills to compound forms that consist of several smaller hillocks. Relief is generally less than 25 feet, (8 m), although in places it may be as much as 60 feet (18 m) and side slopes are variable. These features may be found throughout the morainal belt, but they typically are found along the moraine's inner margin. Collectively, they make up the largest areas in the moraine. These landforms probably represent places where supraglacial debris collected in hollows at the glacier's terminus. Over time the icy substrate melted letting down its sediment load on the land; the thicker areas of sediment now forming the higher parts of the moraine.

Plateaus form flat-topped, broad to slightly arched hills underlain by till. They are absent in the study area, but they have been found farther eastward (Chapter VII, in this guidebook). Many parts of the moraine have a subdued morphology, marked by broad elevated areas of low hummocky relief (Fig. 6). In places these higher areas have ice-contact scarps along their inner borders. Their origin is unclear. They may reflect places where readvancing ice has planed the moraine's surface, or the configuration of stagnant ice and supraglacial debris did not lend itself to forming distinct morainal topography.

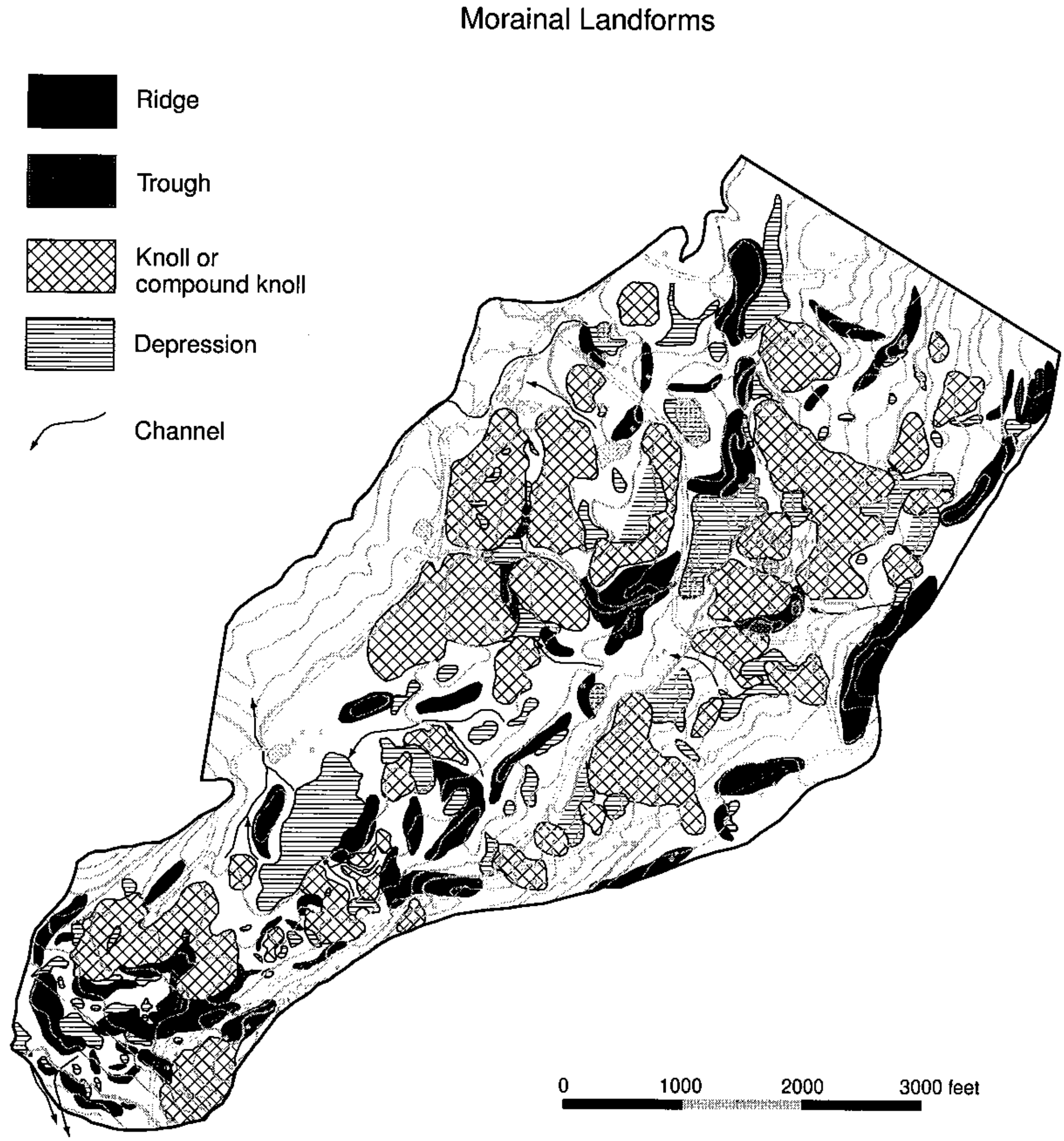


Figure 6. Morphology of the Mt. Mohepinoke segment of the Terminal Moraine, Warren County, New Jersey. Morainal landform elements collectively define areas of ridge-and-trough, and knob-and-kettle topography. Figure from Witte (in press.)

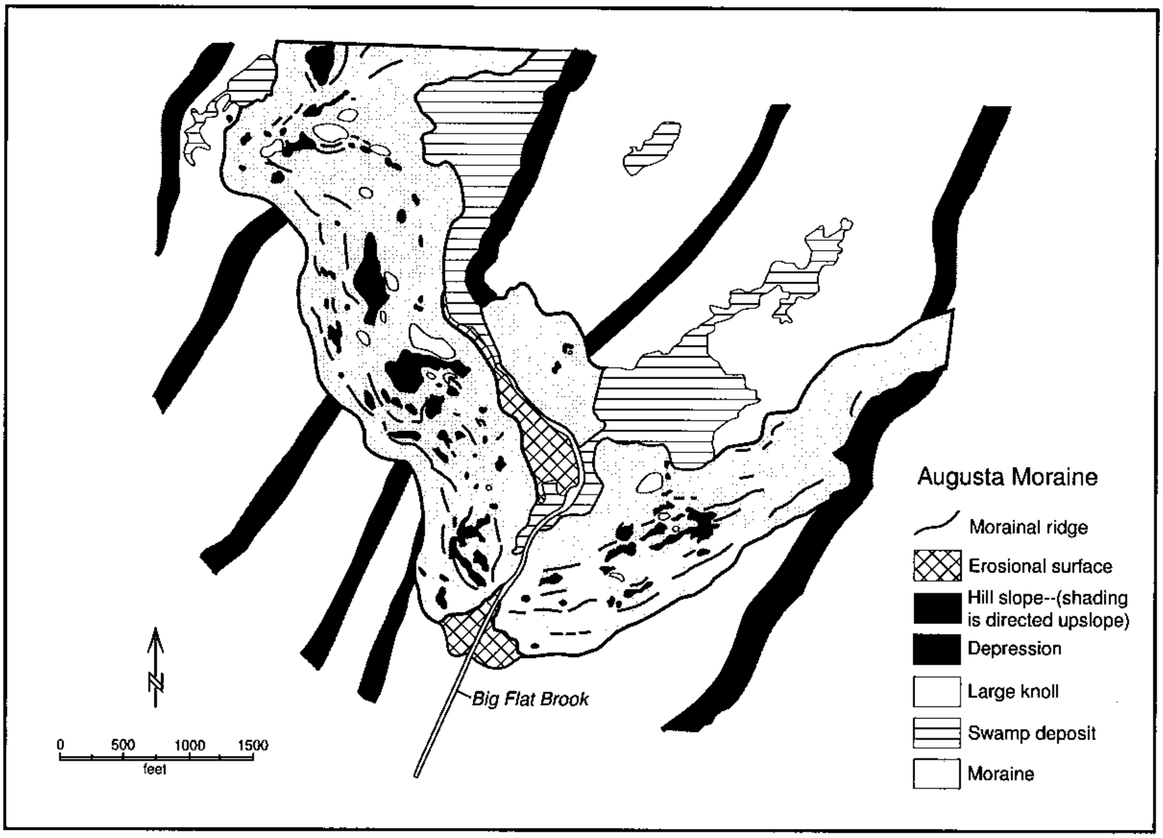


Figure 7. Morphology of the Augusta moraine where it crosses Big Flat Brook valley, Kittatinny Mountain, Sussex County, New Jersey. Morainal landform elements collectively define areas of ridge-and-trough, and knob-and-kettle topography. Figure from Witte (in press.)

Negative topographic elements include troughs, kettles, and meltwater channels. Troughs are elongated depressions that typically parallel MPR's. They are best formed in the outer part of the morainal belt, where in many places they separate nested sets of MPR's (Fig. 7). They are as much as 40 feet (12 m) deep, 100 feet (30 m) wide, and 300 feet (91 m) long. These troughs represent places of little to moderate sediment accumulation between MPR's or they may have originally been ice-cored ridges. Kettles are circular to irregularly-shaped steep-sided depressions. In places they are only partially enclosed forming small amphitheater- shaped bowls. They are as much as 40 feet (12 m) deep, and as much as 500 feet (152 m) wide. Many depressions are wet and contain swamp or bog deposits. Other depressions are dry or contain seasonal water. Kettles have formed where detached blocks of residual ice have melted, leaving behind topographic depressions. In places, low-lying morainal areas are formed by several enclosed to partially enclosed depressions and bowls. They represent the opposite form of the compound morainal knoll and they formed where residual ice initially held up the higher areas along the glacier's margin.

In places the moraine is cut by small, narrow, straight to sinuous channels. These features may be as much as 40 feet deep, and typically have bouldery floors. Today they are used by emphemeral streams and they also carry off the discharge from small springs. Some channels probably formed during the earlier phases of moraine formation when meltwater streams emanating from the active glacier margin flowed along the moraines outer border. Later, during stagnation, meltwater chiefly from melting stagnant ice, drained from hollow to hollow and eventually formed a loosely organized drainage network.

Typically, the innermost parts of the morainal segments have fewer ridges, fewer elongated depressions, and are marked by knob-and-kettle rather than ridge-and-trough topography. The morphology expressed by the Augusta moraine (fig. 7) is typical for morainal segments that abut thick and widespread till. Overall these segments are larger, more continuous, and have more fully developed moraine-parallel ridges than those abutting thin patchy drift. This strongly suggests that unconsolidated material near the glacier's terminus may supply most of the sediment that makes up the moraine rather than glacially eroded bedrock.

Composition

End moraines consist of non compact, bouldery, silty-sandy to sandy till with minor beds and lenses of water-laid sand, silt, and gravel. This material is distinctly different from the more compact, and less stony ground moraine or till that lies near the moraine. Additionally, stratified drift is not a major constituent, even in places where the moraine crosses river valleys or former glacial lake basins. The lithology of the moraine is decidedly local in origin. This was noted by Salisbury (1920, p. 254) who reported that "... the lithologic composition of the till varied from point to point, according to the nature of the formations over which the ice has passed." For example, in the Delaware Valley where the moraine rests on outwash, it contains many rounded, waterworn stones that mimic the provenance of the outwash. Where as on Jenny Jump Mountain, crystalline materials make up the bulk of the moraine. Outcrops of morainal materials are rare due to the difficulty of digging the bouldery drift and more importantly its lack of economic value. The best outcrops are places where the moraine has been removed to expose economic deposits of sand and gravel, such as Saxton Falls and Foul Rift (see Chapter VII in this guidebook). The few exposures observed by the author show that the moraine consists largely of till with minor interlayers and lenses of sorted sand, silt, and gravel. Upon close inspection most of the till is faintly layered with individual layers varying greatly in thickness from less than one foot to as much as 10 feet (3 m). Layering is typically sub horizontal, its base marked by a concentration of larger stones and crude normal grading has been observed in some of these beds. The heterogeneity of morainal sediment, its indistinct layering and grading, and inclusion of water-laid sand, silt, and gravel beds and lenses suggest that most of this material has had a complex history of deposition and is chiefly a product of mass wastage.

Deposits of Glacial Meltwater Streams

Most sediment carried by meltwater streams in Kittatinny Valley was deposited as ice-contact and non-ice-contact deltas, lacustrine-fan deposits, and lake-bottom deposits (Figs. 8 and 9). Smaller quantities of sediment were deposited in valley trains, outwash fans, meltwater-terrace deposits, and in a few kames, ice-channel fillings and eskers.

The distribution of glaciolacustrine deposits in Kittatinny Valley indicates the former existence of many large glacial lakes (Fig. 10). South of the Delaware River-Hudson River drainage divide, all lakes except Lake Newton were dammed by glacial drift downvalley. Lake sluiceways were developed chiefly over these sediment dams. In many places the narrow parts of the lake basins filled completely with deltaic sediment, and were covered by a thick wedge of valley-train sediment that extended from valley wall to valley wall. Glacial lakes in the Wallkill River valley were dammed on the north by the margin of the Kittatinny Valley lobe and they generally discharged southward over bedrock-floored spillways. These lakes expanded northward along the retreating glacier margin until lower outlets, located over local drainage divides in Kittatinny Valley were uncovered, and lake levels dropped or the lakes drained.

Based on the composition of pebbles in outwash and till (Ridge, 1983; Witte, 1988; and Witte and Evenson 1989), most of the outwash was not transported very far from its source. Meltwater streams carried material to the glacier margin through glacial tunnels, as did meltwater streams draining ice-free upland areas beside the valley or lake basin. Sources of sediment include till from beneath the glacier, debris in the basal dirty-ice zone, and eroded till in upland areas. Pebble-count data also suggest that subglacial meltwater streams did not cross drainage divides between adjacent valleys.

In Kittatinny Valley, meltwater streams deposited sediment in fluvial and lacustrine settings (Fig. 9). Glaciofluvial sediments were laid down by meltwater streams in valley-train, meltwater-terrace deposits, and delta topset beds. These sediments include cobbles, pebbles, sand, and minor boulders laid down in channel bars, and sand, silt, and pebbly sand in channel fill and minor overbank deposits. Sediments laid down near the glacier margin in valley-train deposits, and delta-topset beds typically include thick, planar-bedded, and imbricated coarse gravel and sand, and minor channel-fill deposits that consist largely of cross-stratified pebbly sand

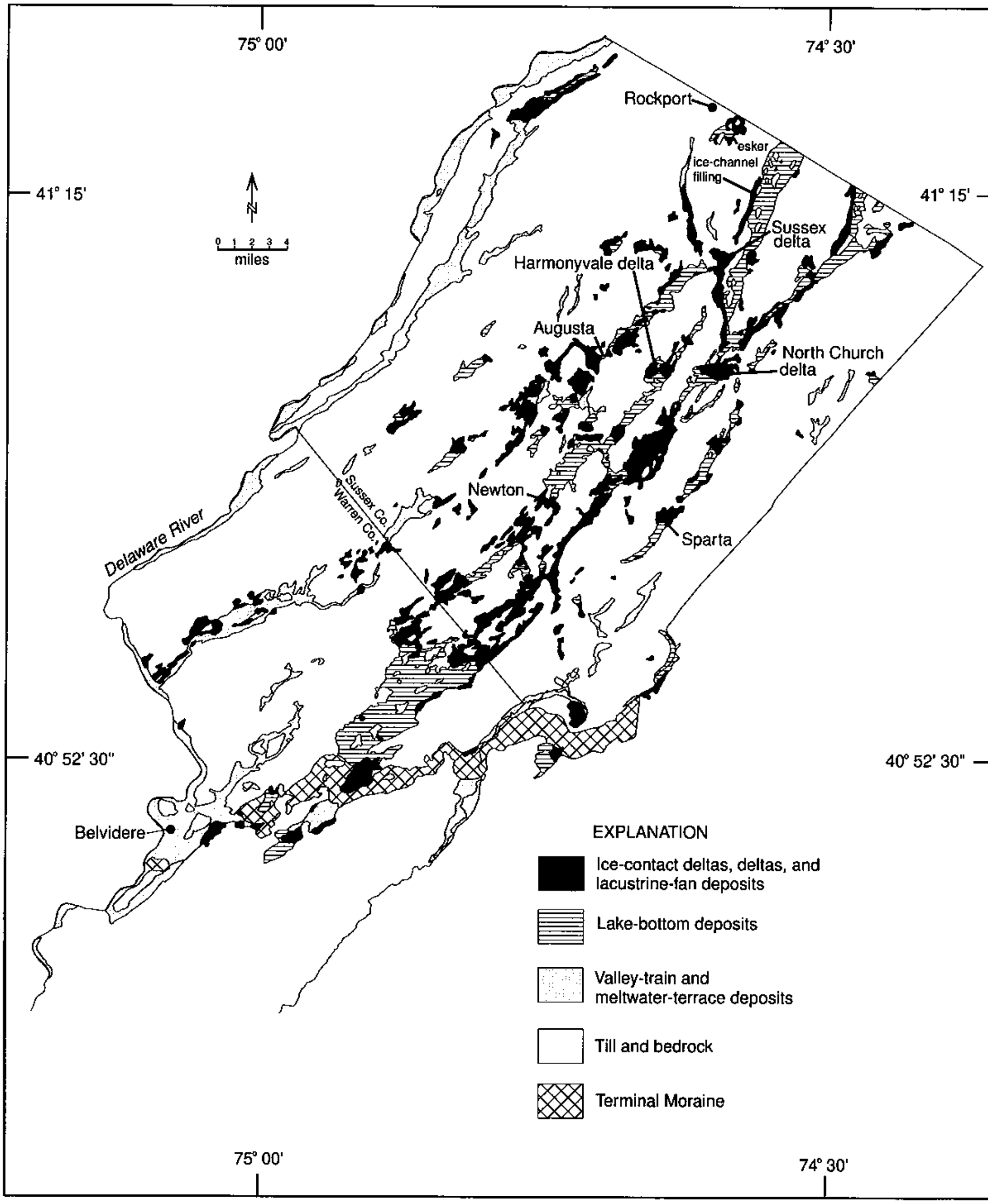
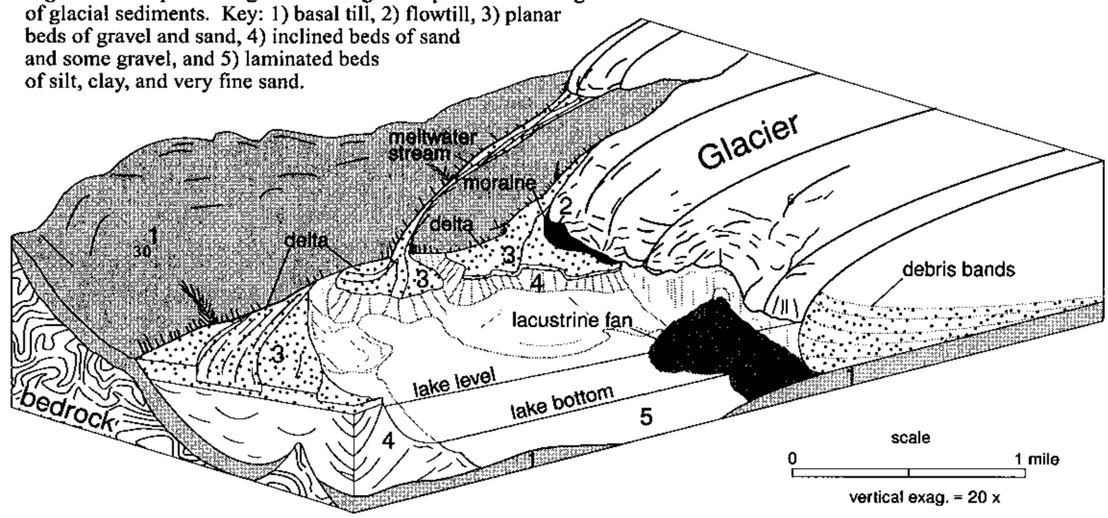


Figure 8. Surficial geologic map showing the late Wisconsinan Terminal Moraine, and meltwater deposits in Sussex and Warren Counties, New Jersey. Alluvial, swamp, and stream-terrace deposits are not shown. Modified from Witte (1997a).

Planar-bedded gravel and sand (deposit of a glacial meltwater stream).

Basal Till - poorly sorted, nonstratified mix of clay- to boulder-sized material.

Figure 9. Composite diagram showing the depositional setting of glacial sediments. Key: 1) basal till, 2) flowtill, 3) planar beds of gravel and sand, 4) inclined beds of sand and some gravel, and 5) laminated beds of silt, clay, and very fine sand.



and sand. Downstream (farther from the glacier's margin), the overall grain size typically decreases, sand is more abundant, and crossbedded and graded beds are more common.

Glaciolacustrine sediments were laid down by meltwater streams in glacial lake deltas, lacustrine-fan deposits, lake-bottom deposits, and in ice-hole fillings mapped as kames. Deltas consist of topset beds of coarse gravel and sand overlying foreset beds of fine gravel and sand. Near the meltwater feeder stream, foreset beds are generally steeply inclined (25° to 35°) and consist of thick to thin, rhythmically-bedded fine gravel and sand. Farther out in the lake basin these sediment's grade into less steeply dipping foreset beds of graded, ripple cross-laminated and parallel-laminated sand and fine gravel with minor silt drapes. These in turn grade into gently dipping bottomset beds of ripple cross-laminated, parallel-laminated sand and silt with clay drapes.

Unlike deltas, lacustrine-fan deposits lack topset beds. They were laid down at the mouth of glacial tunnels that generally exited the glacier near the floor of the lake basin. In Lake Wallkill, a few fans may have also been laid down on the floors of unroofed glacial meltwater tunnels that were connected to the lake. Lacustrine fans also become progressively finer grained basinward. However, near the former tunnel mouth, sediments may be coarser grained and less sorted because of high sedimentation rates and little chance for sorting. If the tunnel remained open and the ice front remained stationary, the fan may have built up to lake level and formed a delta. Sedimentary layering is similar to deltas, except foreset beds deposited near the tunnel mouth are more flat lying, or may dip toward the glacier margin forming backset beds.

Lake-bottom deposits include 1) glacial varves and 2) subaqueous-flow deposits. Glacial varves consist of stacked annual layers that consist of a lower "summer"

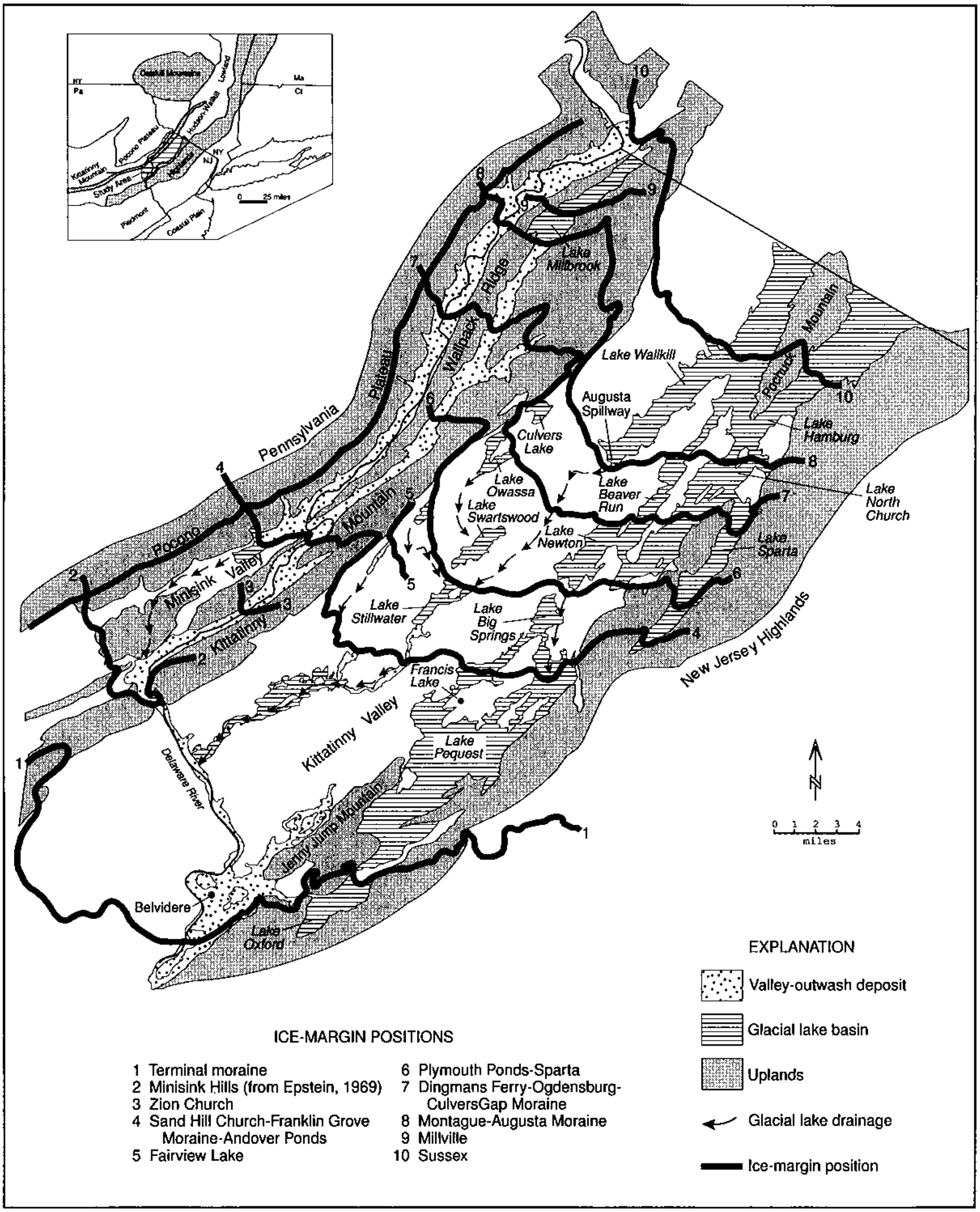


Figure 10. Late Wisconsinan ice-marginal positions of the Kittatinny and Minisink Valley ice lobes, and location of large glacial lakes, extensive valley-outwash deposits in northwestern New Jersey, and northeastern Pennsylvania. Modified from data by Crowl (1971), Epstein (1969), Minard (1961), Ridge (1983), and Witte (1988 and 1997a).

layer of chiefly silt that grades upward into a thinner "winter" layer of very fine silt and clay. Most of these materials were deposited from suspension, although the summer layer may contain sand and silt carried by density currents. Each summer and winter couplet represents one year. Subaqueous-flow deposits consist of graded beds of sand and silt. These deposits originate from higher areas in the lake basin, such as the prodelta front, and are carried down slope into deeper parts of the lake basin by mass flows. Glacial varves grade laterally into bottomset beds of deltas and lacustrine-fan deposits.

Kames consist of a varied mixture of stratified sand, gravel, and silt interlayered with flowtill. In many places they lie above local glacial lake, and topographic base-level controls. However, most exposures revealed collapsed deltaic foreset bedding. Presumably kames were laid down in meltwater ponds that formerly occupied an ice-crevasse, ice-walled sink, or moulin near the glacier's edge.

GEOMETRY of the ICE MARGIN

Because many end moraines in northwestern New Jersey are nearly continuous belts, they define the edge of the ice sheet both geometrically and temporally. The moraines clearly show that the margin of the Laurentide ice sheet was distinctly lobate at both a regional scale and local scale (Figs. 5 and 11). Based on the tracing of end moraines from the valley floor onto adjacent uplands, the surface gradient of the Kittatinny Valley lobe varied between 125 and 290 feet per mile within the first few miles from its margin. In northeastern Pennsylvania, Crowl and Sevon (1980) determined that the slope at the terminus of the Laurentide ice sheet varied from 80 to 405 feet per mile with a "best measure" estimated at 225 feet per mile. The trace of the moraines and estimates on the surface slope of the ice sheet show that extensive lobation as postulated by Ward (1938) did not occur. In fact there exists a similarity between the geometry of the ice margin defined by Terminal Moraine and the recessional moraines located 30 miles away.

Morainal ice margin positions (Fig. 10) are used to construct a chrono-morphostratigraphic framework for the late Wisconsinan deglaciation of northwestern New Jersey. Not only do they provide direct evidence of ice retreat (snapshot), but they also constrain the reconstruction of other ice-retreat positions based on the location of outwash heads, meltwater channels, and glacial lakes. The geometry of ice-retreat positions indicated by the moraines shows that the Laurentide ice sheet retreated in a systematic manner to the northeast, its margin consisting of two distinct sublobes with the largest in Kittatinny Valley and a smaller sublobe in Minisink Valley.

Outwash heads and glacial lakes

Based on the morphosequence concept of Koteff and Pessl (1981), many ice-retreat positions have been mapped in Kittatinny Valley (Ridge 1983; Witte 1988, 1997a) chiefly by identifying outwash heads of ice-contact deltas (Fig. 9). The interpretation of glacial lake histories also provides a basis for reconstructing the ice-retreat history of the Kittatinny Valley ice lobe. For example, ice retreat in the upper part of the Wallkill River valley is marked by ice-contact deltas laid down in Lake Sparta (Fig. 12). As the ice front retreated northward in the valley, lower spillways were uncovered to the west along a drainage divide between the Paulins Kill and Wallkill River, and the elevations of the deltas become correspondingly lower. These intrabasin relationships occur throughout the many drainage basins in Kittatinny Valley, and they show that regional or valley-ice lobe stagnation is not a valid style of deglaciation.

In Kittatinny Valley the correlation of coeval ice-retreat positions is based on: 1) lobate ice-margin geometry, 2) tracing of moraines, 3) location and relative size of outwash deposits laid down along stable ice margins, 4) location of meltwater channels in uplands, and 5) shape of the glacier margin required to dam lakes or to cover lower lake outlets. Based on these five constraints five major ice-recessional positions have been recognized in the upper part of Kittatinny Valley. They are the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex margins (Fig. 10). Each position is traceable across the valley; their course marked by end moraines and/or ice-contact deltas, and they form the basis from which the deglacial history of the Kittatinny Valley lobe can be reconstructed. Many minor recessional positions have been delineated in Kittatinny Valley by Ridge (1983), and Witte (1988, 1991, and 1997a). These positions are generally marked by ice-contact deltas and they provide additional evidence that deglaciation was systematic and chiefly characterized by stagnation-zone retreat. However, these ice-recessional margins are untraceable across Kittatinny Valley. Therefore, they are not discussed within the regional framework of deglaciation presented here.

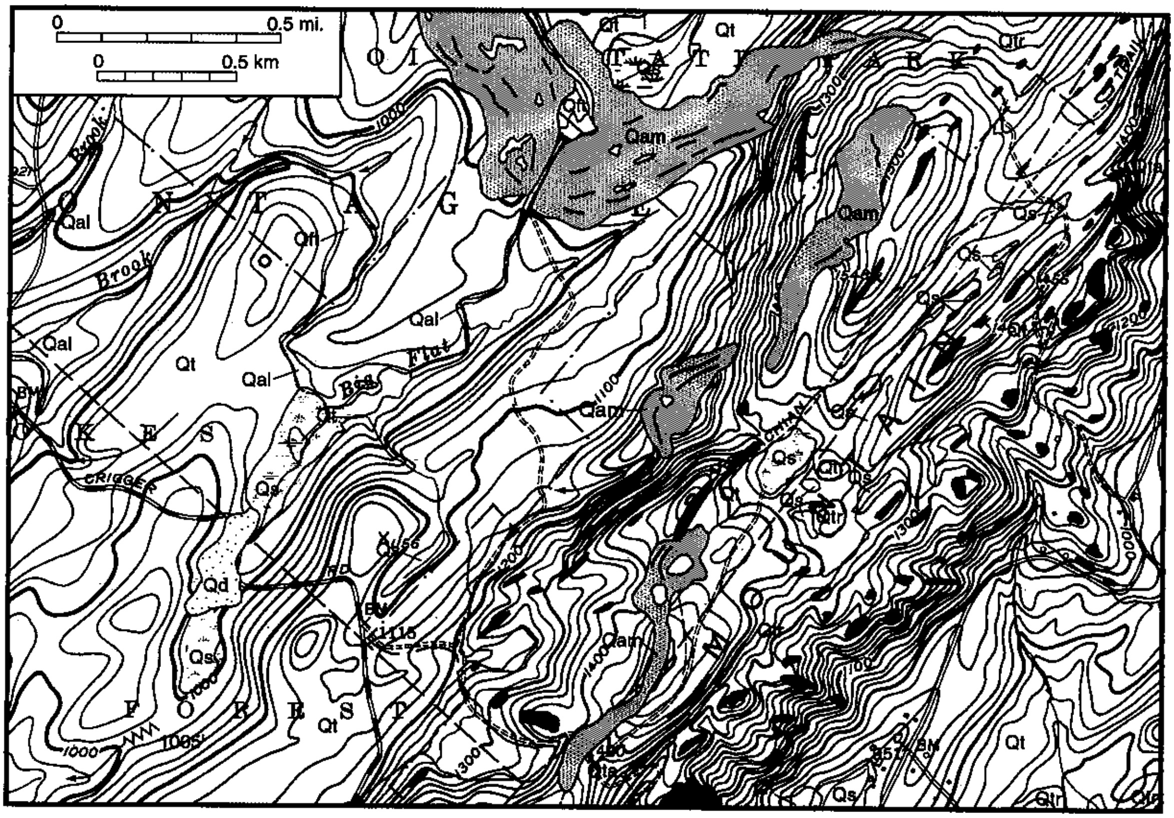


Figure 11. Surficial geologic map of part of the Branchville 7 1/2 minute topographic quadrangle. The area shown is located on Kittatinny Mountain in the NW 1/4 of the quadrangle. List of map units: Qal - alluvium, Qs - swamp and bog deposits, Qta - talus, Qt - thick till, Qtr - thin till, Qam - Augusta Moraine, Qd - ice-contact delta. The unlabeled, dark gray areas represent extensive bedrock outcrop and very thin till. The heavy black lines shown on the moraine represent the crest of moraine-parallel ridges, and the small unlabeled white areas are depressions. Figure modified from Witte (1997a).

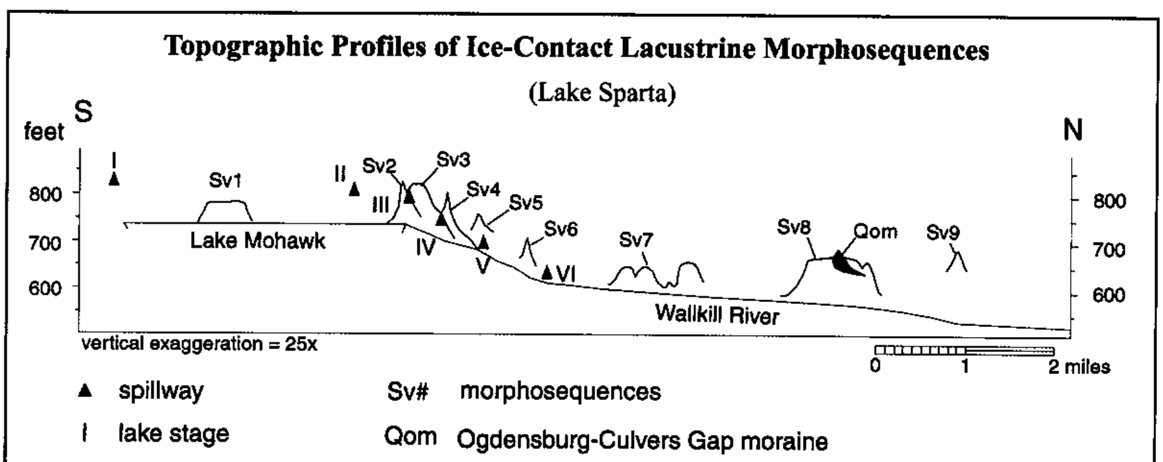


Figure 12. Topographic profiles of 9 ice-contact-lacustrine morphosequences in the upper part of the Wallkill Valley, Sussex County, New Jersey. Morphosequences consist of a lacustrine-fan deposit (Sv1) laid down in stage I, and ice-contact deltas Sv2 and Sv3 laid down in stage II, Sv4 in stage III, Sv5 in stage IV, Sv6 in stage V, and Sv7, Sv8, and Sv9 in stage VI of Lake Sparta. During deglaciation, the level of Lake Sparta lowered as the glacier margin retreated northward, and lower spillways were uncovered on a drainage divide between the Wallkill River and Paulins Kill. The corresponding decrease in delta elevations indicates deglaciation was systematic, and not the result of regional stagnation. Figure modified from Witte (1997a).

Other Indicators of Ice-retreat Positions

Additional evidence that shows that the margin of the Kittatinny Valley lobe was lobate includes striations and the location of meltwater channels. The orientation of striations in Kittatinny Valley (Fig. 13) records a pattern of divergent flow at the glacier's margin. Lines of ice flow indicated by younger striations (those formed during deglaciation) are perpendicular to ice margin positions based on the tracing of end moraines. This ice flow pattern could have only developed at or near a lobate margin where thicker ice in the valley's center spread outward to thinner areas along the valley walls. Divergent striations also show that ice was active near the glacier's margin. In places meltwater channels also mark the glacier's margin. These channels typically parallel nearby morainal margins and were probably cut by meltwater streams flowing along the glacier's margin from uplands toward river valleys. They are best formed in areas underlain by thick till where bedrock does not influence the stream's course.

FORMATION of END MORAINES: ACTIVE and STAGNANT ICE at the GLACIER MARGIN

The study of an end moraine's morphology, course, and composition show that they are complex landforms, their genesis not easily described by simplistic depositional models. The character of the moraine shows that both active ice and stagnation play a role in their formation. The following definition, modified from Flint (1971) adequately describes the character of these features. "An end moraine is a ridge-like accumulation of drift built along any part of the margin of an active glacier. Its topography is initially constructional, and its initial form results from (1) amount and vertical distribution of drift in the glacier, (2) rate of ice movement, and (3) rate of ablation." Flint stressed the role of active ice transporting drift to the glacier margin, and the amount of drift in the ice sheet. Presumably, the more active the glacier and the more drift it contains, the larger the end moraine it will make. In addition, syndepositional and postdepositional modification of the moraine through ice shove, collapse due to melting of buried ice, and resedimentation of supramorainal materials chiefly by mass wastage, all act to give the recessional moraines their overall form. Figure 14 shows through a time lapse series of panels how end moraines may have formed. Based on their course, morphology, and composition, end moraines formed under the following set of conditions:

1) Active ice must be present to transport glacial debris to the glacier's terminus. This requires that the glacier must be temperate (warm-based) and that part of its forward movement occurs by basal sliding.

2) There must be a substantial zone of basal debris (basal dirty ice) at the glacier's sole. Sediment, chiefly contained in debris bands, is carried upward at the glacier's terminus by compressive flow, shearing, and folding, where at its surface, sediment is released from ice by melting. Supraglacial sediment is further transported by mass movement, running water, and in some instances ice shove.

3) Because the transport of glacial debris is slow, the formation of thick end moraine requires that the glacier's margin must remain static (neither advancing nor retreating) or that it only oscillates throughout a narrow marginal zone (x 10^2 to x 10^3 feet). Length of the stillstand is estimated at one thousand to fifteen hundred years for the Terminal Moraine and several hundred years for the larger recessional moraines. These estimates are based on a late Wisconsinan retreat history of about four thousand years for New Jersey (Chapter I in this guidebook).

4) Accumulation of debris across the glacier's terminus is variable due to variations in basal debris content and rates of basal sliding at the glacier's bed. This coupled with differences in ice thickness, due to local topographic relief, results in differential melting and the creation of a more uneven supraglacial surface.

5) Eventually the debris-covered terminus becomes a margin of stagnant ice with the leading edge of active ice moving back up the glacier. As melting proceeds in the stagnant zone, supraglacial debris is slowly let down on the land, the former supraglacial topography becomes inverted with sediment filled basins forming the higher areas and high-standing ice blocks or ice-cored ridges forming the low areas.

6) Minor oscillations of the glacier margin can redistribute sediment, override stagnant ice, and leave additional stagnant blocks during the post-readvance phase of melting.

End moraines require active ice to form, yet their overall morphology is the result of stagnation and redistribution of sediment by mass wastage. The lobate course of the moraines, their morphology, and evidence of glacial readvance suggests they were formed by 1) the transport

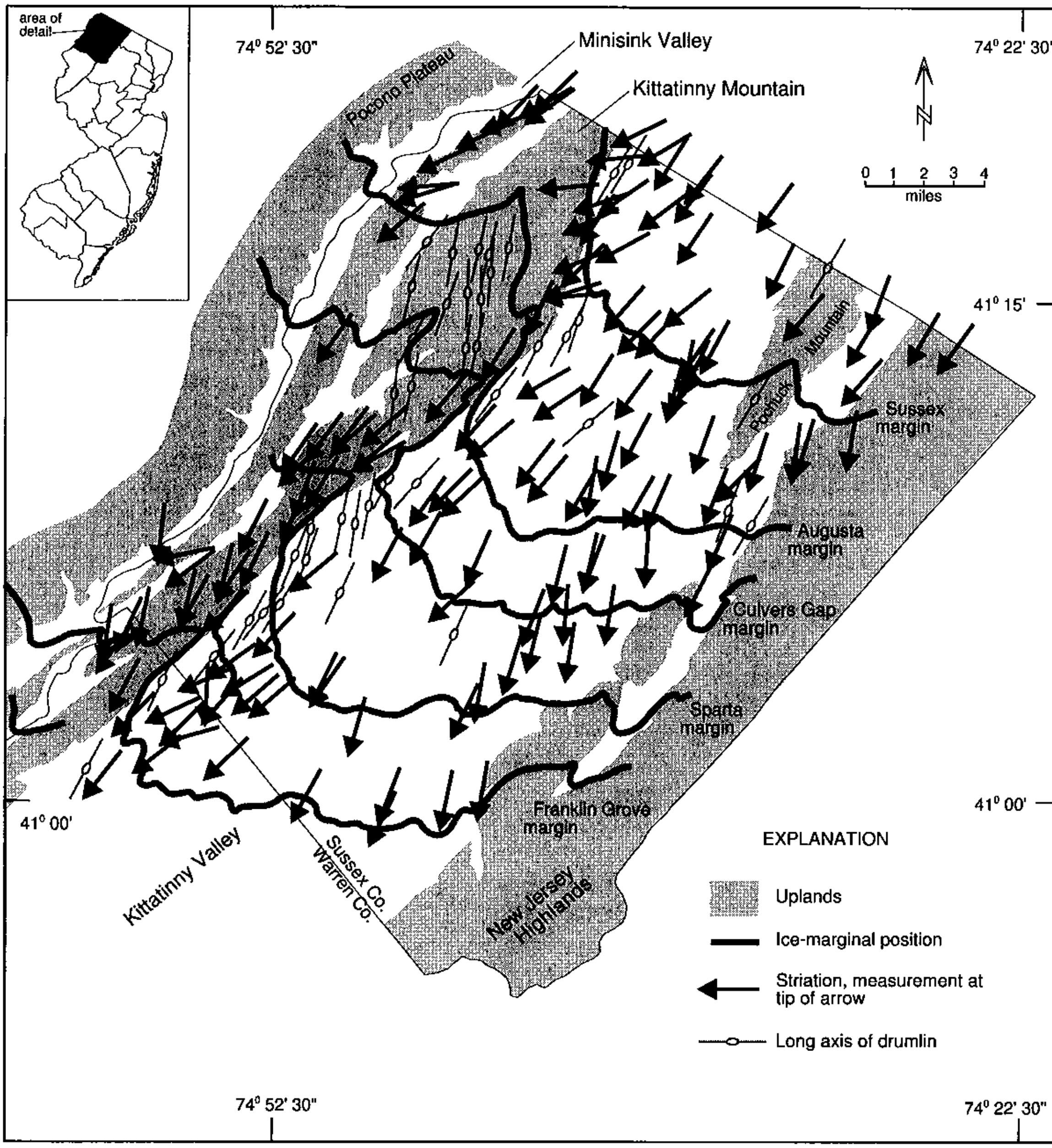


Figure 13. Orientation of striae, and drumlins in the upper part of Kittatinny Valley and surrounding area in Sussex and Warren Counties, New Jersey. Figure modified from Witte (1997a).

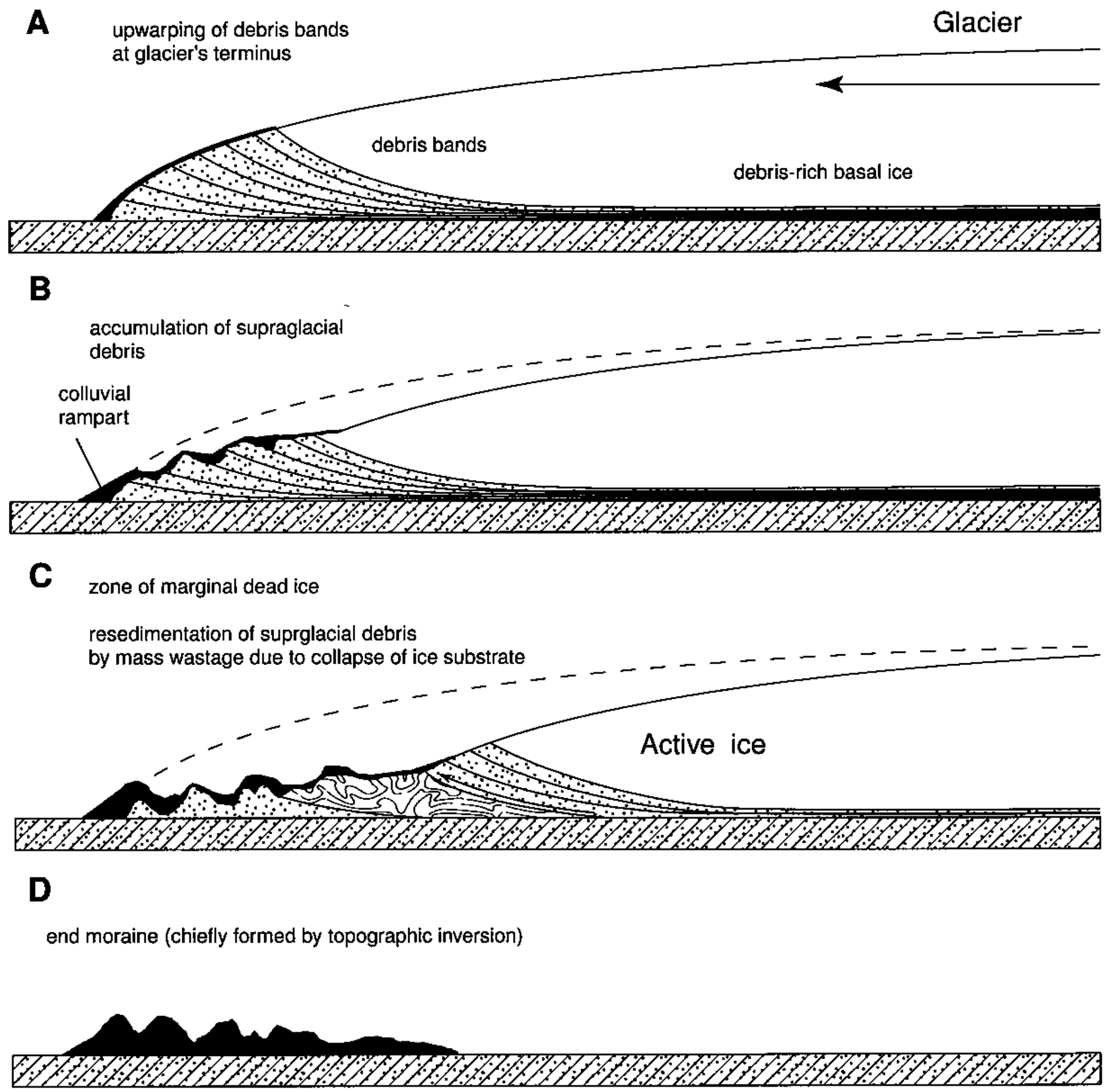


Figure 14. A sequence of panels showing how end moraines may have formed in New Jersey. Panel A - Active ice is present at the glacier's terminus. Sediment, chiefly basal debris transported along debris bands, is carried upward at the glacier's terminus by compressive flow, shearing, and folding of the debris-rich basal ice. Panel B - Sediment, released from ice by melting, accumulates along the glacier's margin where it collects in hollows and crevasses. Because the amount of debris is also variable, differential melting occurs, further increasing relief. Over time and if the glacier's terminus remains in a relatively constant position, a debris blanket of varying thickness covers the marginal area of the glacier. Panel C - Contiuned accumulation of debris and differential melting causes the terminal area of the glacier to become very thin. This results in stagnation and the formation of a marginal zone of dead ice. Because the leading edge of active ice shifts backwards the transport of debris to the glacier's former active ice margin has ceased. Panel D - Gradually the suprglacial debris is let down on the land by the continued melting of dead ice and resedimentation by mass wastage. Except for colluvial ramparts and push ridges, morainal morphology is largely a product of topographic inversion where high areas of glacial ice covered by a veneer of dedris now form the low areas (kettles and hollows) and low areas filled with a thick accumulation of debris now form the high areas (some morainal ridges, knolls and hillocks). Minor oscillations of the glacier's margin and several closely-spaced stillstands may redistribute sediment, override stagnant ice, and leave additional stagnant ice forming a complex assemblage of morainal landforms. Figure modified from Flint (1971, Fig. 5-14).

of debris and debris-rich ice by the glacier at its margin, and 2) penecontemporaneous and postdepositional sorting and mixing of material by mass movement, chiefly resulting from slope failure caused by melting ice, and saturation and collapse of sediment. The source and mechanism of sediment transport is unclear. Most of the morainal material is of local origin, but it is not known whether the glacier was reworking drift at its margin or transporting sediment to its margin by direct glacial action. Inwash is not a viable mechanism because the larger deposits lie on mountain or ridge tops. The origin of the moraine-parallel ridges is also unclear. These features may be push ridges. However, due to the absence of data on their internal structure, this premise is highly speculative.

GLACIAL HISTORY

Glacial Advance and Changes in Direction of Regional Ice Flow

The late Wisconsinan history of ice advance into Kittatinny Valley is ambiguous because glacial drift and striae that record this history has been eroded or were buried. If the ice sheet advanced in lobes as suggested by the lobate course of the Terminal Moraine, then its initial advance was marked by lobes of ice moving down the Kittatinny and Minisink Valleys. Sevon and others (1975) suggested that ice from the Ontario basin first advanced southward into northeastern Pennsylvania and northwestern New Jersey. Later, ice from the Hudson-Wallkill lowland, which initially had lagged behind, overrode Ontario ice. In this scenario the course of the Terminal Moraine in Minisink and Kittatinny Valleys was controlled by ice flowing from the Hudson-Wallkill lowland. Connally and Sirkin (1973) suggested that the Ogdensburg-Culvers Gap moraine represents or nearly represents the terminal late Wisconsinan position of the Hudson-Champlain lobe based on changes in ice flow noted by Salisbury (1902) near the moraine. Ridge (1983) proposed that a sublobe of ice from the Ontario basin overrode Kittatinny Mountain and flowed southward into Kittatinny Valley. Southwestward flow occurred only near the glacier margin where ice was thinner, and its flow was constrained by the southwesterly trend of the valley (Fig. 15a). Analyzes of striae, drumlins, and the distribution of erratics in the upper part of Kittatinny Valley and adjacent Kittatinny Mountain support Ridge's view. However, a Ontario or Hudson-Champlain source cannot be solely assigned to ice in Kittatinny Valley based on the evidence of southerly ice flow in the upper part of the valley (Fig. 15a). These data further show that by the time the Ogdensburg-Culvers Gap moraine was formed, ice flow in Kittatinny Valley had turned completely to the southwest with extensive lobation at the margin (Fig. 15b).

Striae in the southern part of Kittatinny Valley and on Kittatinny Mountain south of Culvers Gap show that glacier flow was southward there. From the Terminal Moraine northeastward up through Kittatinny Valley, striae show a stronger valley-controlled lobate flow pattern. A few sets of cross-cutting striae near the Franklin Grove moraine (Witte 1988) show ice flow shifting to the southwest in Kittatinny Valley, and striae on Kittatinny Mountain north of the Augusta margin show a westerly flow direction. This suggests that the glacier's margin was very lobate, and that active ice was present at or very near the ice lobe's margin. The orientation of drumlins in Kittatinny Valley and on Kittatinny Mountain (Fig. 5) also indicates that ice initially flowed southward over Kittatinny Mountain into Kittatinny Valley, and was also later superseded by ice flowing to the southwest.

The distribution of erratics shows a similar change in ice flow. Boulders of quartz-pebble conglomerate derived from the Shawangunk Formation, and boulders of red sandstone derived from the Bloomsburg Formation, both of which makes up Kittatinny Mountain, are dispersed throughout Kittatinny Valley and the western part of the New Jersey Highlands. In some places near the base of Kittatinny Mountain, they make up as much as 80 percent of the gravel-sized fraction in till (Ridge, 1983; Witte, 1988; Witte and Evenson 1989).

Nephelene syenite boulders in glacial drift of Kittatinny Valley were first recognized by Salisbury (1902, fig. 29, p. 106). Their distribution shows that the direction of ice flow in the valley ranged from S 12° W to S 46° W. North of Culvers Gap boulders of syenite lie on Kittatinny Mountain, and a few are due west and as much as 500 feet (152 m) above the syenite outcrop body. Boulders of sandstone from the Martinsburg Formation in Kittatinny Valley are also on Kittatinny Mountain north of Culvers Gap. The location of these erratics here indicates extensive lobation, and divergent flow at the margin of the Kittatinny Valley lobe.

Boulders of nephelene syenite and Martinsburg hornfels, uncovered by a recent landslide in Minisink Valley, provide additional information on the advance history of the Kittatinny and Minisink Valley ice lobes. The boulders were apparently derived from till that had covered a

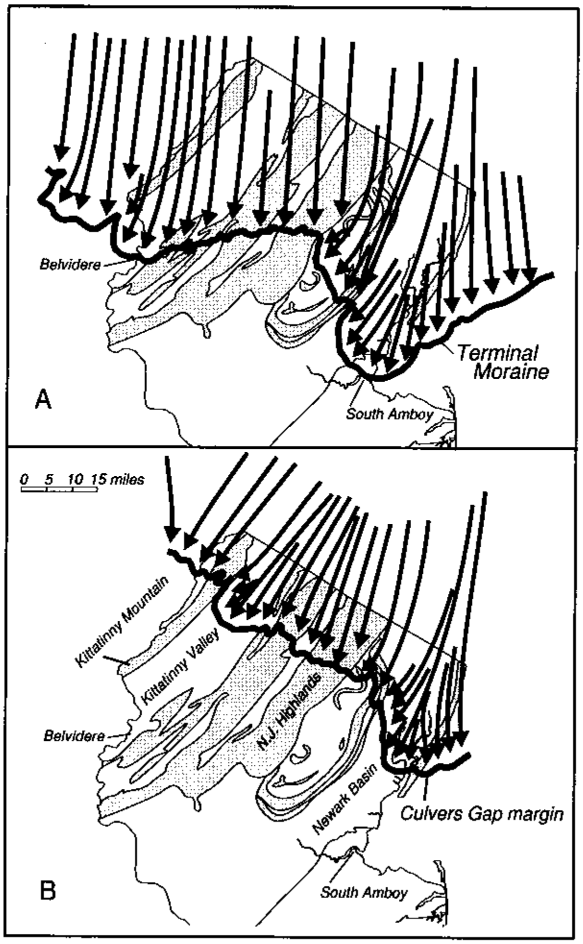


Figure 15. Generalized direction of ice movement in northern New Jersey during the late Wisconsinan. Lines represent regional ice-flow movement at the base of the ice sheet. Flow directions are based on striae, drumlins, dispersal of erratics, and till provenance. Shaded areas represent major uplands. Figure 15a shows direction of ice flow when the glacier margin was at the Terminal Moraine. Field data in the Kittatinny Valley area indicates ice flowed southward across the valley's southwest-trending regional topographic grain. Figure 15b shows direction of ice flow during deglaciation. Flow lines in Kittatinny and MinisinkValleys and surrounding uplands are oriented in a southwest direction with well developed lobate ice flow at the glaciers margin. The change in regional ice flow to a southwest direction appears to be related to thinning of the ice sheet at its margin, and reorganization of ice flow around the Catskill Mountains, and in the Hudson-Wallkill Valley. Data from Ridge (1983), Stanford and Harper (1985), Witte (1988), Sevon and others (1989), Stone and others (1989), and unpublished field maps on file at the New Jersey Geological Survey, Trenton, New Jersey.

steeply-dipping bedrock slope. Previous investigations by Salisbury (1902), Ridge (1983), and Witte (1988, 1991) have shown that the boulders lie in Kittatinny Valley, and on Kittatinny Mountain northward of Culvers Gap. Based on their location in Minisink Valley, ice would have had to flow S 55^0 west in a direction across Kittatinny Mountain, from the outcrops in Kittatinny Valley. This direction is not consistent with the direction of ice flow reconstructed from striae, drumlins, and till provenance elsewhere in the area. A possible explanation for the location of the syenite-bearing till may be that it represents the initial advance of late Wisconsinan ice into this area from the Hudson-Wallkill lowland. Later as the ice thickened and the Kittatinny and Minisink Valleys lobes coalesced, ice flow turned southward.

The distribution of gneiss erratics in Kittatinny Valley also confirms the shift in ice flow from south to southwest. In the southern part of the valley they are absent. A few erratics are found near the Ogdensburg-Culvers Gap moraine, and become more abundant to the northeast in the upper part of the Wallkill Valley. Most are derived from the metasedimentary rocks from Pochuck Mountain, and they further corroborate the development of a well-defined valley-flow system by the time the Ogdensburg-Culvers Gap moraine was deposited.

The initial advance of the late Wisconsinan ice sheet into the lower part of Kittatinny Valley was marked by the southward movement of ice over Kittatinny Mountain into the valley (Fig. 15a). On reaching Jenny Jump Mountain the Kittatinny Valley lobe split into the smaller Delaware Valley and Pequest Valley sublobes. These continued to advance southward with the thickening glacial ice gradually burying the northern parts of Jenny Jump Mountain. During the maximum extent of the ice sheet only a small part of the mountain's southern end was not covered by ice. In front of the advancing ice sheet proglacial outwash was laid down in the Delaware and Pequest Valleys. Some of this material may be observed in a sand and gravel pit near Foul Rift in the Delaware Valley (Chapter VII, in this guidebook) where it lies beneath basal till. Continued advance of the Kittatinny Valley lobe down the Delaware Valley resulted in the formation of glacial Lake Oxford (Figs. 16 and 17) when ice advanced into the western end of Pequest Gap and dammed the valley. The lake rose in level until a low drainage divide between Pophandusing Brook and the Pequest River, near the site of the Oxford Stone Quarry (585 feet (179 m)), was reached and the lake spilled out into the Pophandusing Brook valley. Promorainal ice-contact deltas in the Pequest Valley south of Townsbury, mark the terminal position of the Pequest Valley sublobe, and provide evidence of the former existence of glacial Lake Oxford. The lake and its deposits were described earlier by Ridge (1983).

The limit of the late Wisconsinan glacial border is well defined in Kittatinny Valley and nearby uplands, as it is elsewhere in New Jersey. It is either marked by the outer edge of the Terminal Moraine, or as it is in most places by till that extends as much as 3000 feet (914 m) out beyond the moraine. In western New Jersey the glacial border is defined by the most southern occurrence of fresh dolostone till stones on the surface. The distribution of late Wisconsinan till and ice-contact meltwater deposits out beyond the Terminal Moraine is consistent with the stratigraphy observed in the Delaware River valley near Foul Rift (Chapter VII, in this guidebook), which shows that the Delaware Valley sublobe extended southward from the position marked by the Terminal Moraine. Dense basal till overlying proglacial outwash marks the advance of the ice sheet to a position south of the Terminal moraine.

Following advance to its most southerly position, the Kittatinny Valley lobe retreated northward of the position marked by the Terminal Moraine. This retreat is evidenced by occurrence of stratified materials overlying basal till in the Foul Rift section (Chapter VII, in this guidebook). The extent of retreat is unknown. However, based on the thickness of outwash laid down during this stage, the Delaware Valley sublobe presumably retreated to a position just north of the position marked by the Terminal moraine and stayed there a short time. Following this short-lived retreat, the glacier readvanced and deposited the Terminal Moraine. In the Delaware Valley it is locally called the Foul Rift moraine. Morainal segments elsewhere have been named Bridgeville, Jenny Jump, Mountain Lake, Mt. Mohepinoke, and Townsbury. Regionally, the Terminal Moraine is a chronostratigraphic unit that marks or very nearly marks the late Wisconsinan glacial border. Locally it may represent as much as fifteen hundred years of deposition at the margin of a very slowly retreating ice lobe. Based on the geometry and morphology of the Mt. Mohepinoke segment (Fig. 6), the outer part of the moraine is older than its inner part. The width of the morainal belt near Moores Pond compared with the much narrower moraine downvalley, and the tracing of moraine parallel ridges suggest that in places the Terminal Moraine was laid down from several ice retreat positions as previously suggested by Ridge (1983). The late Wisconsinan border in front of the Terminal moraine in the Delaware and Pequest Valleys also suggests that the oscillation of the glacier margin may have been more pronounced in the valleys rather than the mountains due to the effects of high elevation and relief on glacial flow.

A large ice-contact delta between the Foul Rift and Bridgeville moraines, called here the Pophandusing delta (Fig. 16), appears coeval with the moraines based on its location. The delta rises to an elevation of 485 feet (148 m) and it appears to have been laid down in a large ice-walled pond that formed between the eastern edge of the Delaware Valley sublobe and a large bedrock ridge on the western flank of the New Jersey Highlands. A spillway on a drainage divide between Pophadusing and Buckhorn Creeks appears to have controlled the elevation of the delta. Foreset beds measured in the northern part of the deposit dip southeastward and they suggest the delta may have been laid down at the mouth of a subglacial tunnel in the Delaware Valley.

At some time before or synchronous with the formation of the Terminal Moraine was the lowering of Lake Oxford to a level now marked by deltaic deposits that are as high as 530 feet (161 m) (Fig. 17). These deposits are correlative with the Townsbury segment of the Terminal Moraine. This lake may have spilled out through a threshold located near St. Nicholas Church (525 feet (160 m), or it may have followed a course along the edge of the Mountain Lake and Bridgeville sublobes.

Style and Timing of Deglaciation

The deglacial history of the Laurentide ice sheet is well documented for northwestern New Jersey and parts of eastern Pennsylvania. Epstein (1969), Ridge, (1983), Cotter and others (1986), Stone and others (1989), and Witte (1988, 1997a, in press) showed that the margin of the Kittatinny Valley and Minisink Valley lobes retreated in a systematic manner with minimal stagnation. However, the age of the Terminal Moraine, timing of the late Wisconsinan maximum, and precise chronology of deglaciation are very uncertain. This is due to scant radiocarbon dates because of a lack of organic material that can be used to date deglaciation, inadequacies of dating bog-bottom organic material and concretions, and use of sedimentation rates to extrapolate bog-bottom radiocarbon dates. Also, there are few exposures of varves that can be used for chronology.

The few radiocarbon dates available bracket the age of the Terminal Moraine and retreat of ice from New Jersey. Radiocarbon dating of basal organic material cored from Budd Lake by Harmon (1960) yielded a date of 22,890 +/- 720 yr B.P. (I-2845), and a concretion sampled from sediments of Lake Passaic by Reimer (1984) that yielded a date of 20,180 +/- 500 yr B.P. (QC-1304) suggests that the age of the Terminal Moraine is about 22,000 to 20,000 yr B.P. Basal organic material cored from a bog on the

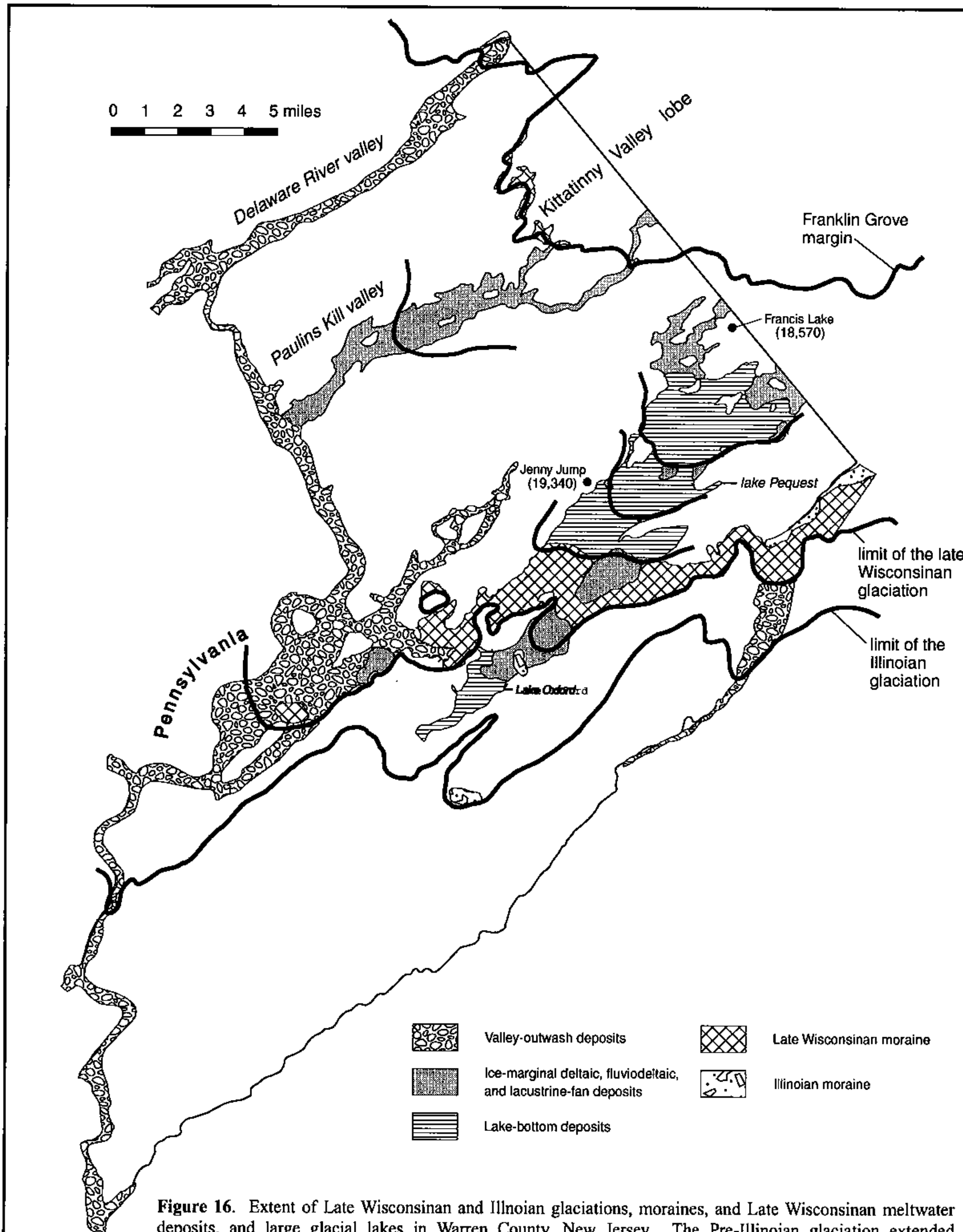


Figure 16. Extent of Late Wisconsinan and Illnoian glaciations, moraines, and Late Wisconsinan meltwater deposits, and large glacial lakes in Warren County, New Jersey. The Pre-Illinoian glaciation extended southward of the area shown in the figure. Glacial data modified from Ridge (1983), Witte and Stanford (1995), and Stone and others (in press). The Francis Lake date is from Cotter (1983) and the Jenny Jump date is from D.H. Cadwell (written commun., 1996).

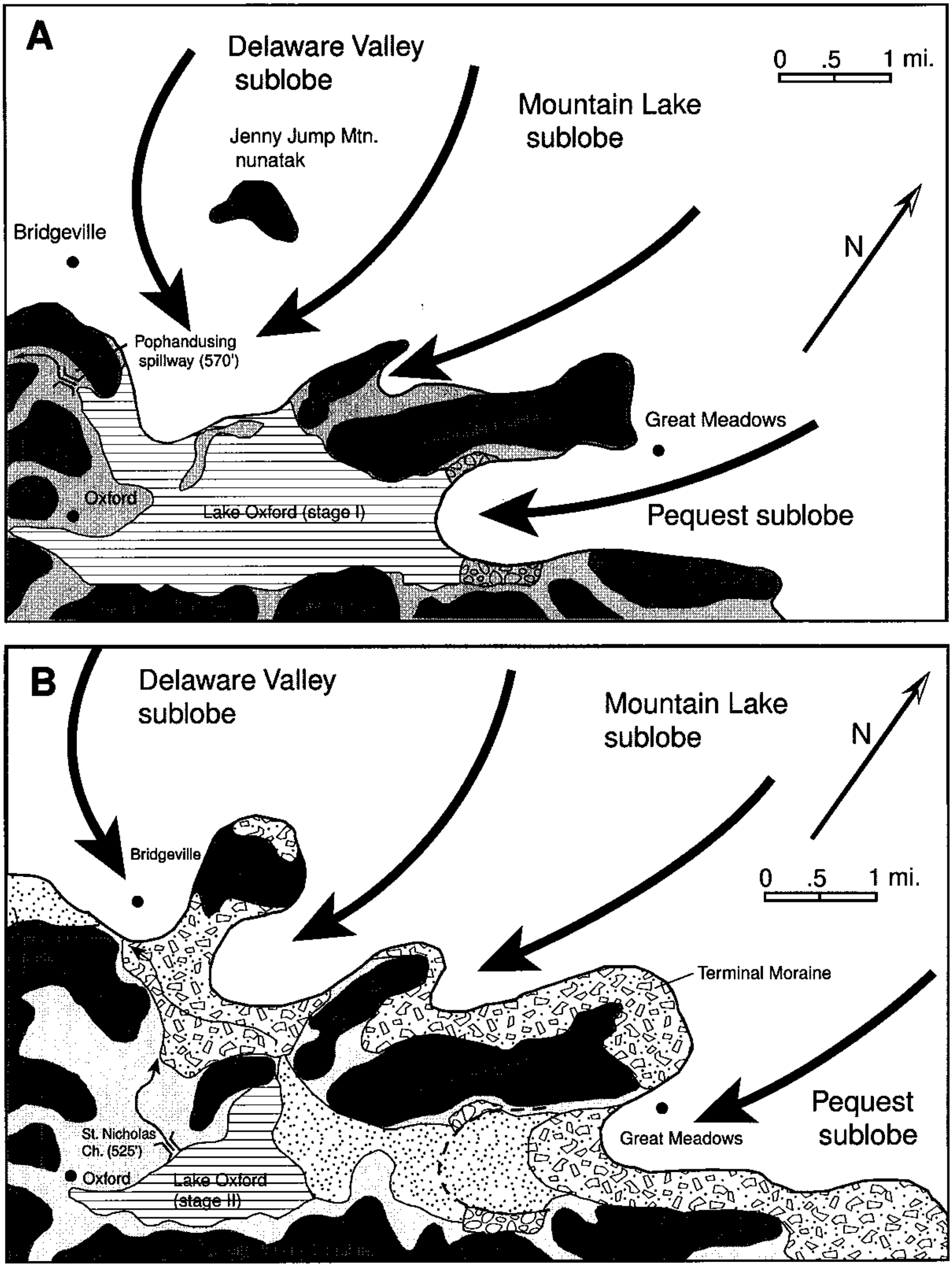


Figure 17. Deployment of ice and geometry of Lake Oxford in the lower part of the Pequest Valley during the late Wisconsinan maximum (panel A) and during deposition of the Terminal Moraine (panel B). Figure modified from Ridge (1983).

side of Jenny Jump Mountain near the Terminal Moraine by D. H. Cadwell (written communication, 1996) and basal organic material cored from Francis Lake by Cotter (1983) indicates a minimum age of deglaciation at 19,340 +/- 695 yr B.P. (GX-4279), and 18,570 +/- 250 yr B.P. (SI-5273) respectfully for the lower part of Kittatinny Valley. Because Francis lake lies about 3 miles (5 km) southeast of the Franklin Grove moraine, this age is also used here as a minimum date for that feature. Exactly when the ice margin retreated out of the New Jersey part of Kittatinny Valley is also uncertain. A concretion date of 17,950 +/- 620 yr B.P. (I-4935) from sediments of Lake Hudson (Stone and Borns, 1986) and an estimated age of 17,210 yr B.P. for the Wallkill moraine by Connally and Sirkin (1973) suggest ice had retreated from New Jersey by 17,500 yr B.P.

Because the Terminal Moraine and recessional moraines in northwestern New Jersey were formed at the margin of an active ice sheet, the ice margins they define may not accurately reflect the geometry of the ice lobe if a significant amount of stagnant ice existed beyond the active glacier. Stagnant ice or "dead" ice is glacier ice that is no longer flowing forward. Stagnant ice may consist of a valley sublobe many miles long (Ward, 1938; Crowl, 1971), or small detached blocks left by a retreating glacier. A margin of dead ice may have also bordered on the glacier's active margin, more or less synchronously wasting back with the retreating active glacier margin. This style of retreat, called *stagnation-zone retreat*, was originally defined by Currier (1941) and modified by Koteff and Pessl (1981) to describe deglaciation in New England (largely defined by the outwash heads of ice contact deltas laid down in proglacial lakes) where end moraines were not found. This style of deglaciation was proposed by Ridge (1983), and Witte (1988), for Kittatinny Valley and later modified by Witte (1991, and 1997a) to account for an active glacier margin and recessional moraines.

Retreat from the Terminal Moraine and deglaciation in the lower part of Kittatinny Valley

Delaware Valley

Retreat from the Terminal Moraine position resulted in the abandonment of the Pophandusing delta as the glacier margin retreated off the east wall of the Delaware Valley. Ice-contact deltas laid down behind the Bridgeville moraine define minor recessional positions of the retreating glacier margin (Fig. 18). The highest of these at 460 feet (140 m) appears to be on grade with a meltwater channel cut down in the Pophandusing delta, which lies at 450 feet (137 m). Small deposits of recessional moraine east of Bridgeville and a small ice contact delta north of Belvidere delineate minor retreatal positions of the Kittatinny Valley lobe and define the geometry of the Beaver Brook and Delaware Valley sublobes. Based on the occurrence of fines beneath coarser outwash south of Belvidere it appears that the Foul Rift moraine may have dammed the Delaware Valley upon retreat from the Terminal Moraine position. Outwash supplied from the erosion of the Bridgeville and Mountain Lakes moraines by meltwater from glacial Lake Oxford and ice-retreat positions farther north filled in this sediment-dammed lake basin. Furthermore, meltwater terrace deposits cut down in this outwash, show that the moraine dam was slowly eroded over time lowering local base-level control.

North of the Bridgeville area no other retreatal positions have been identified in the valley. Outwash from retreatal positions farther north and erosion by meltwater from glacial Lake Pequest in the Pequest Valley continued to carve lower terraces into higher deposits along the course of the Delaware and Pequest Rivers.

Lower Pequest Valley

Meltwater deposits in the Pequest Valley include ice-contact deltas, lacustrine-fan deposits, and lake-bottom deposits laid down in Lake Pequest. The history of the lake was described earlier by Salisbury (1902), Ridge (1983), and Witte (1988). Lake Pequest (Fig. 16) was a large and deep lake dammed down valley by the Terminal Moraine near the village of Townsbury. There are several sluiceways through the moraine that were probably cut by the lakes's outlet waters. The highest of these lies at 565 feet (172 m) above sea level, and it appears to have controlled the elevation of ice-contact deltas during the early period of the lake's history. At some point this spillway was abandoned and a lower one, apparently located along the course of the Pequest River became active. The timing of this event is not clear, although the elevation of deltas in the northern part of the basin near the village of Johnsonburg (these lie at an elevation of 560 feet, 171 m) suggests that level of the lake had dropped by the time the glacier had retreated to a position in the northern part of the lake basin.

Lacustrine-fan deposits in the valley define two ice-retreat positions of the Kittatinny Valley lobe (Fig. 16). These are called the Post Island and Stevens Island margins. These positions have been traced westward by Ridge (1983) into the Paulins Kill valley where they are

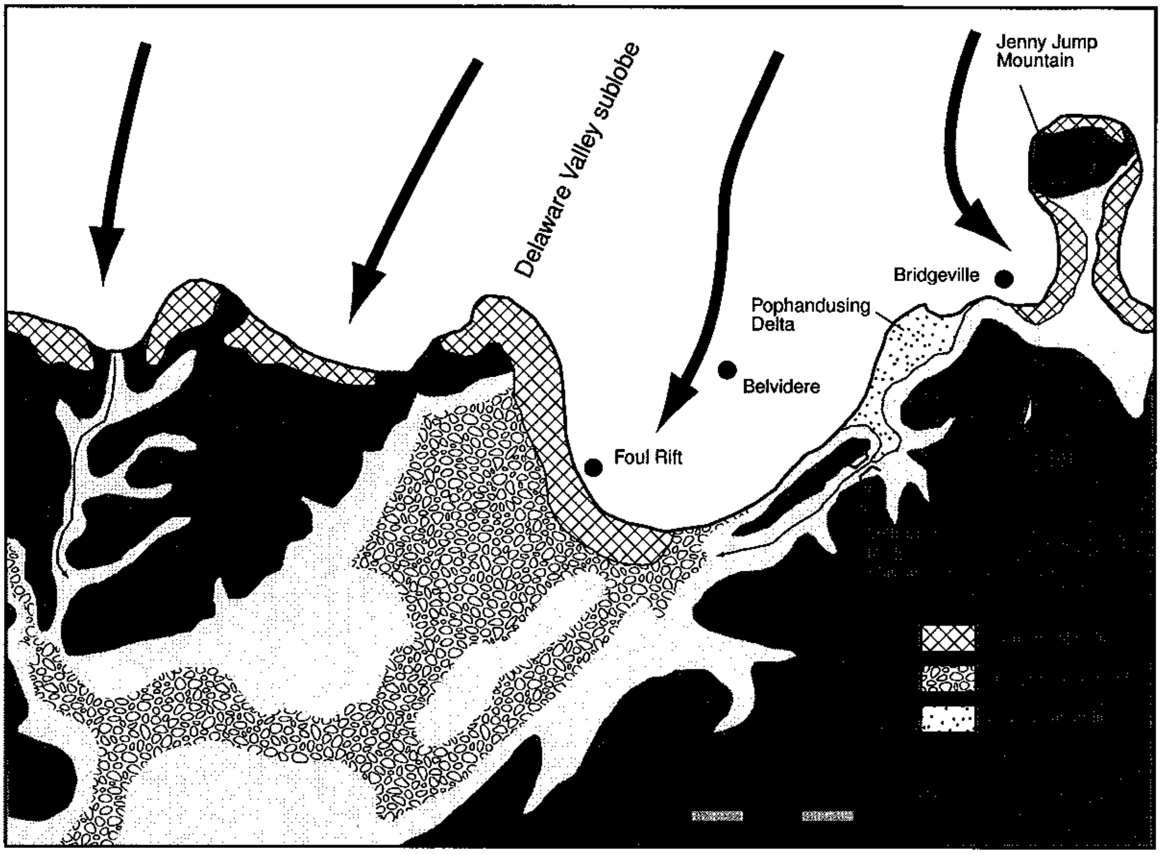


Figure 18. Geometry of the Kittatinny Valley ice lobe during deposition of the Terminal Moraine and Pophandusing delta. Figure modified from Ridge (1983).

tentatively correlated with ice contact deltas near Hainesburg and Walnut Valley. Based on the morphosequence concept of Koteff and Pessl (1981), Ridge (1983) and Witte (1988, and 1991) have delineated 13 ice-retreat positions in the Pequest Valley.

Paulins Kill Valley

Meltwater deposits in the Paulins Kill valley consist of ice-contact, and non-ice-contact deltas, and meltwater-terrace deposits (Fig. 16). These deposits lie as much as 100 feet (30 m) above the Paulins Kill. Deltaic deposits form the bulk of the stratified sediment in the Paulins Kill valley and these deposits were laid down in small proglacial lakes held in the south-draining valley by older outwash deposits downvalley and possibly ice from the Delaware Valley sublobe. The deglacial sequence illustrated in Figure 19 was repeated often in the Paulins Kill valley and based on the morphosequence concept of Koteff and Pessl (1981); Ridge (1983) and Witte (1988) have delineated 14 ice-retreat positions in the valley. The meltwater-terrace deposits that cover part of the valley floor (Fig. 16) were formed by meltwater emanating from up valley positions. The lower positions of the meltwater terraces reflect a lowering of local base level as older outwash down valley became further incised by meltwater draining from younger retreat positions upstream. Meltwater continued to flow down the Paulins Kill valley well after the valley had been deglaciated. Outlet water from Lake Wallkill (Witte, 1997a) continued to flow across the spillway at Augusta and erode meltwater deposits in Paulins Kill Valley, until a lower spillway was uncovered in the Wallkill Valley, and the lake drained into the Hudson Valley.

Upper part of Kittatinny Valley: a change in ice retreat

The retreat history in the upper part of Kittatinny Valley is different to that of its lower part where the margin of the Kittatinny Valley lobe appears to have retreated rapidly to the position marked by the Franklin Grove moraine. Other than the Terminal Moraine, ice-reces-

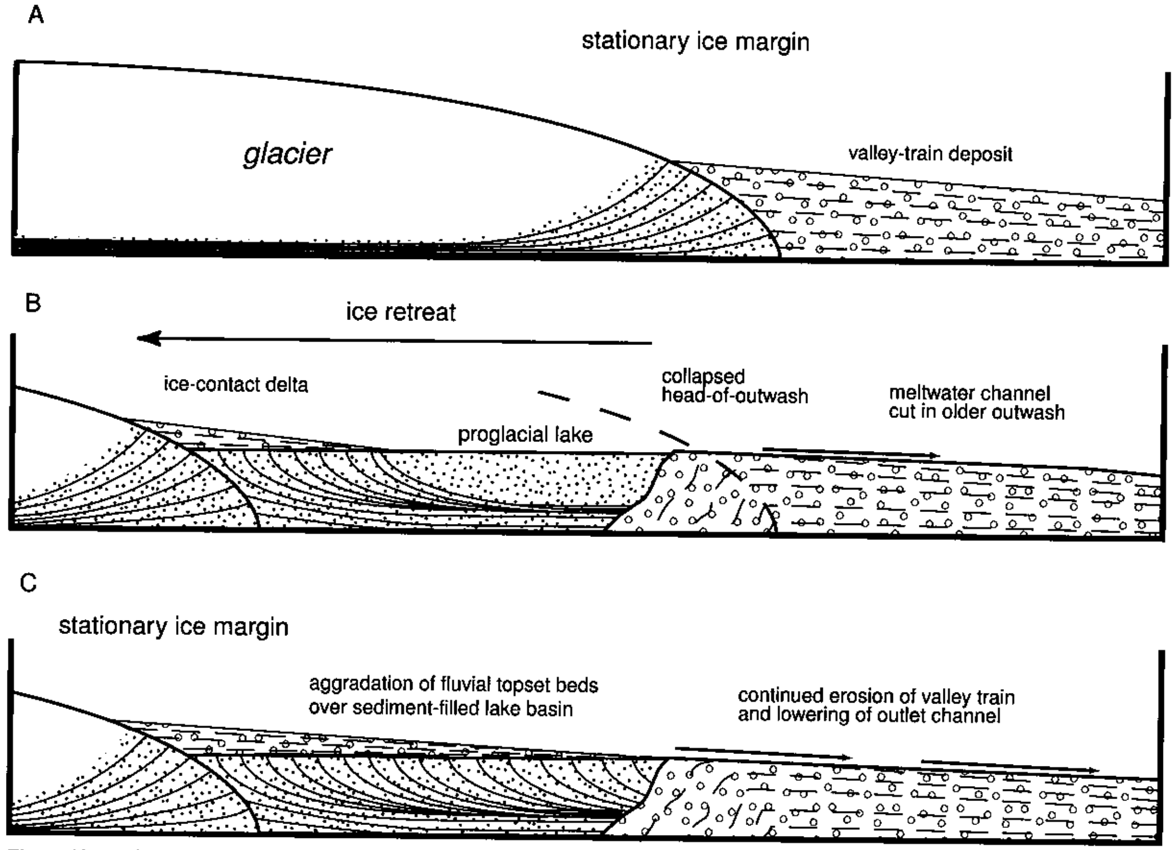


Figure 19. Typical deglaciation sequence in narrow valley draining away from the glacier. In panel A, a large valley train has been deposited at and beyond a stationary ice margin. In panel B, the glacier retreats several miles up valley to where it becomes stationary once again. During retreat from its previous position a proglacial lake has formed between the retreating glacier and the valley train's head-of-outwash. Into this lake an ice-contact delta has been deposited. Over time, the lake's outlet waters erode a channel in the underlying valley train. In panel C, continued deposition in the lake fills the lake's basin with deltaic sediment (chiefly sand and fine gravel). Over time meltwater streams will deposit a thick wedge of coarse gravel and sand burying the ice-contact delta.

sional positions are not readily traced across the lower part of the valley, although deglaciation here was systematic and marked by stagnation-zone retreat. Apparently, the locations of outwash heads were controlled to a very large extent by topography. The difference in the pattern of glacial retreat between the lower and upper parts of Kittatinny Valley marks a change to a more pronounced lobation of the ice lobe during deglaciation, and suggests that other factors besides local topographic control have influenced the retreat history of the Kittatinny Valley lobe.

Five ice margins, the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex have been identified that mark major recessional positions of the Kittatinny Valley lobe. The Sparta, Culvers Gap, Augusta, and Sussex margins are traceable onto the New Jersey Highlands, and all margins, except Sparta, are traceable westward into Minisink Valley. All appear to represent halts in ice retreat, and some mark minor readvances of the Kittatinny Valley lobe. This pattern of ice retreat is different from the more rapid style of retreat postulated for the lower part of Kittatinny Valley. These differences as well as the close spacing of the Culvers Gap and Augusta margins, their large traceable extent, and correlation with extensive end moraines and ice-contact deltas in western New Jersey may indicate that other factors, besides local topographic control, have influenced the retreat history of the Kittatinny Valley lobe. Although the O^{18} record shows that the climate remained cold and stable during the few thousand years that it took for ice to retreat from New Jersey, there exists minor fluctuations that may have influenced deglaciation.

The Franklin Grove margin delineates the first major ice-retreat position that is traceable across Kittatinny Valley (Fig. 20). It is marked by an end moraine and promorainal outwash in Paulins Kill valley. To the east, correlative ice-contact deltas were laid down in Lake Pequest and the initial phase of Lake Big Springs and they delineate the ice margin there. Retreat of the ice margin from the Franklin Grove position was followed by the formation of several proglacial lakes that expanded northward with the retreat of the ice lobe. These are Lakes Stillwater, Big Springs, and the initial phase of Lake Sparta (Fig. 20). Lake Stillwater was dammed downstream by slightly older outwash that had filled a narrow part of the Paulins Kill valley. Similarly, Lake Big Springs was dammed by outwash previously laid down in narrow valleys in the northern part of Lake Pequest, and Lake Sparta formed in the north-draining Wallkill River valley. The Sparta margin (Fig. 20) delineates a second major ice-recessional position and it is delineated by large ice-contact deltas laid down in the aforementioned lakes.

Retreat of the ice lobe to the Culvers Gap margin was marked by the northward expansion of Lake Sparta, and the formation of Lakes Newton, Swartswood, and Owassa (Fig. 20). Lake Sparta consisted of 6 stages (Fig. 12), each corresponding to a decline in lake level as lower outlets, located west of the lake on a drainage divide between the Wallkill River and Paulins Kill, were uncovered by the retreating glacier. The level of Lake Newton was initially controlled by a sluiceway cut in deltas laid down in Lake Big Springs. Just before the Ogdensburg-Culvers Gap moraine was deposited, erosion had lowered the outlet to a position controlled by a bedrock floor (Witte 1988). Lakes Swartswood and Owassa were dammed by till at the south ends of their respective lake basins.

In Kittatinny Valley, the Culvers Gap margin is associated with ice-contact deltas that both predate and postdate the Ogdensburg-Culvers Gap moraine (Witte 1988). Initially, the deltas were laid down in proglacial lakes that occupied parts of the Wallkill River and Paulins Kill valleys. Most of these deltas are large, presumably indicating the stagnant glacier margin remained in the same place for a lengthy period. During a minor readvance of the Kittatinny Valley lobe, the glacier overran parts of these deltas and formed a moraine. Records of wells in Paulins Kill and Wallkill River valleys (Witte 1991) show morainal deposits overlying outwash, and support a readvance of the Kittatinny Valley lobe. Ice-contact deltas and lacustrine fans laid down on the north side of the moraine mark local ice-retreatal positions. These deltaic and morainal deposits had been mapped previously as kame moraine (Spencer and others 1908; Herpers 1961). In Minisink Valley the Culvers Gap margin is also delineated by the Dingmans Ferry moraine (Fig. 5), along with the heads-of-outwash of large valley-train deposits in both the Flat Brook and Delaware River valleys.

Retreat of the Kittatinny Valley lobe from the Culvers Gap margin to the Augusta margin was marked by the drainage of Lakes Sparta and Newton, and the formation of Lakes North Church (Stanford and Harper 1985) and Beaver Run in the upper part of the Wallkill River valley (Fig. 21). Lake Sparta drained when the ice margin uncovered the north end of the Pimple Hills. Lake Newton (Fig. 21), which initially discharged into the Pequest River valley over a bedrock-floored outlet at the south end of its basin, drained when moraine and deltaic outwash, at the north end of its basin near Lafayette, became eroded and the lake drained into the Paulins Kill valley. Meltwater-terrace deposits cut down in Lake Newton deposits on the east side of the lake basin indicate that erosion of the sediment dam had commenced prior to ice retreat from the Augusta margin. Both Lakes Beaver Run and North Church discharged over sluiceways located on a drainage divide between the Paulins Kill and Wallkill River. The former over a bedrock-floored spillway, and the latter over deltaic deposits that had been laid down in Lake Newton.

The Augusta moraine marks a major ice margin and based on the record of a well drilled on it in Papakating Creek valley (Witte, 1997a) it was also deposited following a readvance. The moraine is traceable over Kittatinny Mountain into Minisink Valley where it is called the Montague moraine (Fig. 5). Eastward it is correlative with the Harmonyvale and North Church deltas, which were laid down in Lakes Beaver Run and North Church, respectively.

Retreat of the Kittatinny Valley lobe from the Augusta margin resulted in the initial formation of Lake Wallkill (Witte 1988, in press) in Papakating Creek valley (Fig. 21), and the expansion of Lake North Church in the Wallkill River valley. Although, Connally and others (1989) refer to the former water body as Lake Fairchild, the New Jersey Geological Survey has adopted the name Wallkill based on their standard of naming the largest glacial lakes after the master stream in the lake basin. Initially, Lake Wallkill's spillway was over the Augusta moraine. Eventually the sluiceway was lowered by fluvial erosion into the underlying gravel and sand of an ice-contact delta that had previously filled the Paulins Kill valley south of the position now marked by the moraine. Erosion continued until bedrock was reached, and the level of the lake stabilized. The present elevation of

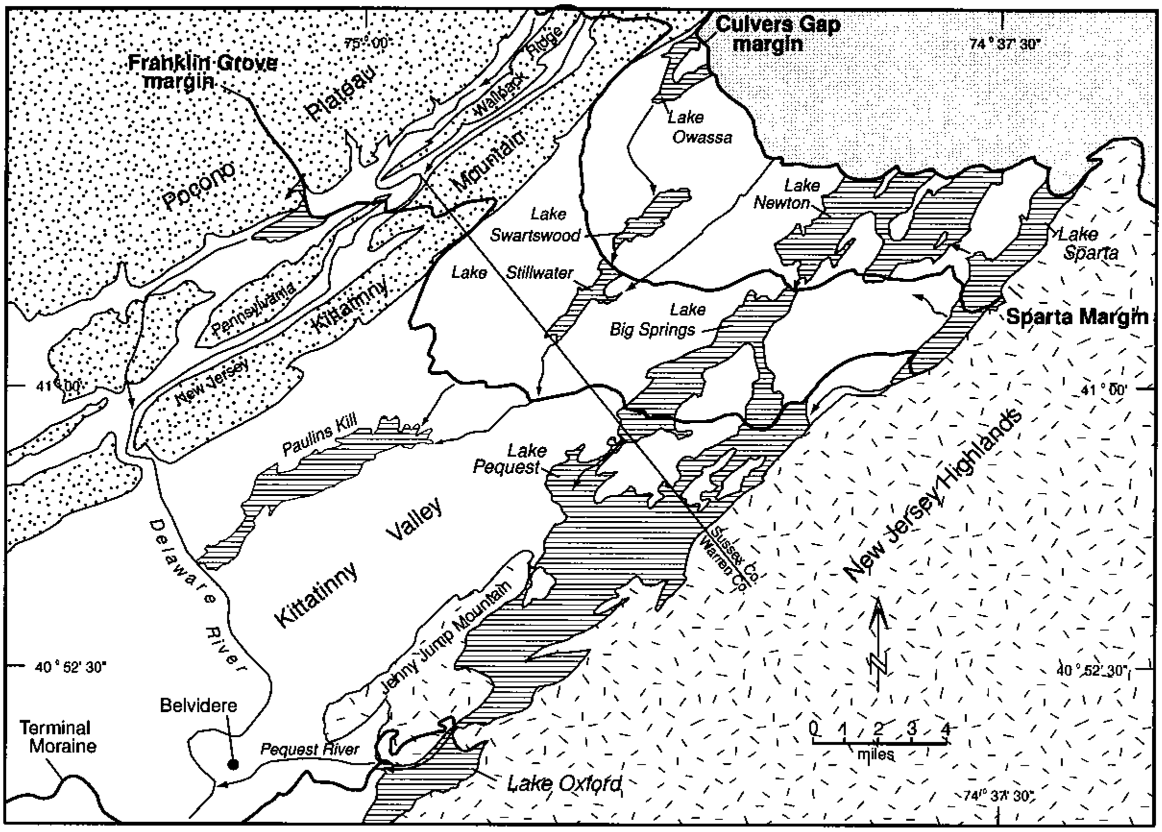


Figure 20. Late Wisconsinan ice-retreat positions, large glacial lakes, and meltwater drainage in Kittatinny and Minisink Valleys, New Jersey and Pennsylvania. Figure modified from Witte (1997a).

this threshold (Fig. 21), located south of the moraine and called here the Augusta spillway, is estimated to be 495 feet (151 m) above sea level. The interval antedating the formation of the Augusta stage is here called the Frankford Plains phase of Lake Wallkill. Based on the elevation of topset-foreset contacts of deltas built into the lake, this phase lasted until the Kittatinny Valley lobe retreated from the Sussex margin (Witte in press).

The next major ice-retreatal position in the upper part of Kittatinny Valley was the Sussex margin, and it is delineated by a large ice-contact delta near Sussex, a small end moraine near Libertyville, and smaller ice-contact deltas in Lake Hamburg (Stanford and Harper 1985). Retreat to this position was accompanied by the lowering of Lake North Church and Lake Beaver Run to the level of Lake Hamburg, which discharged over a local drainage divide into Lake Wallkill (Fig. 21). After the ice margin retreated north of the confluence of Papakating Creek and the Wallkill River, Lake Hamburg lowered to the level of Lake Wallkill.

Lake Wallkill continued to expand northward until ice uncovered the northern end of the Skunnemunk Mountains, and a lower outlet, that now lies at an elevation of 365 feet (111 m) above sea level, was uncovered on a drainage divide between the Wallkill River and Moodna Creek (Adams 1934; Connally and others, 1989). Minor retreatal positions are marked by lacustrine-fan deposits in the main part of the Wallkill River valley, and ice-contact and non-ice-contact deltas along the former shoreline of the lake. At this time the Augusta spillway was abandoned, and in the upper part of the Wallkill River valley thin stream-terrace deposits were laid down on the newly exposed floor of Lake Wallkill. Later the former lake basin became tilted due to isostatic rebound (Koteff and Larsen, 1989), and a shallow lake flooded the upper part of the valley in postglacial time.

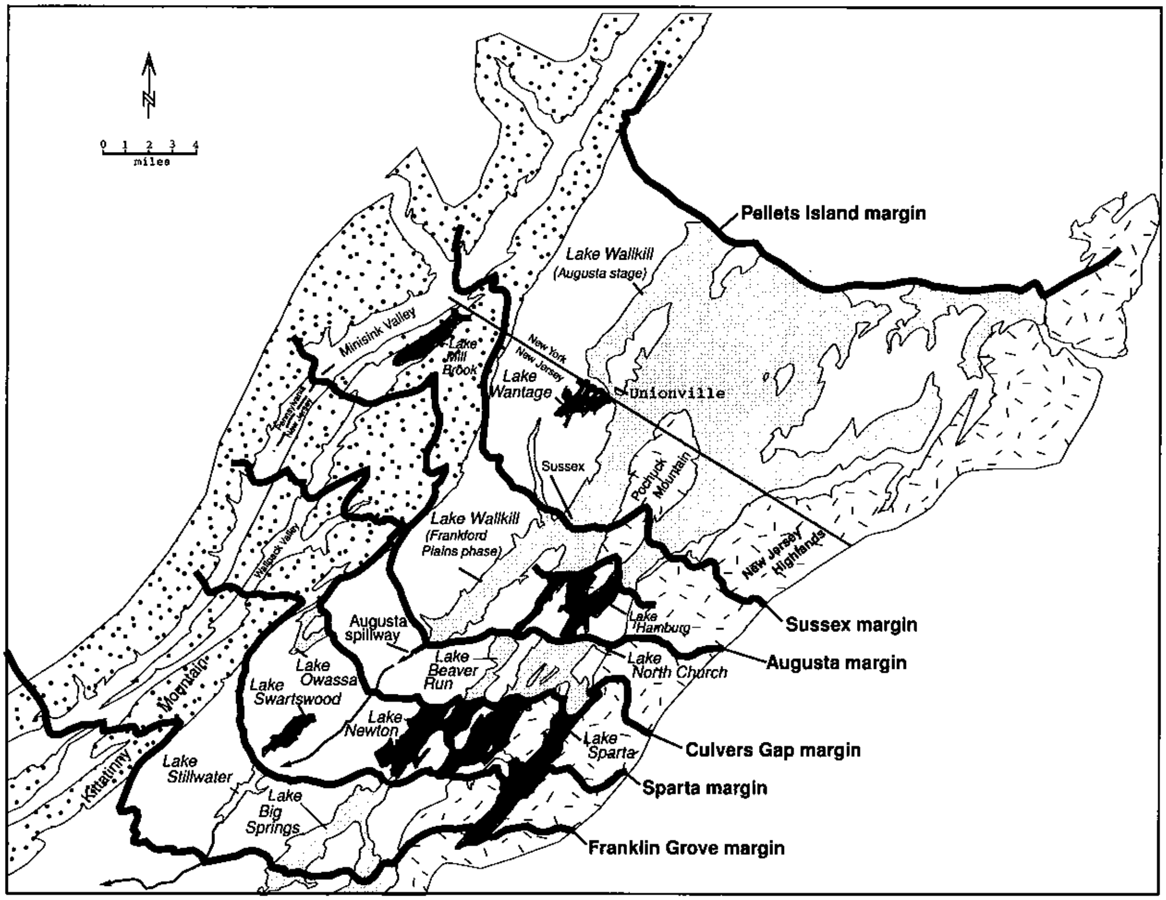


Figure 21. Late Wisconsinan ice-recession margins and glacial lakes in the upper parts of Kittatinny and Wallkill Valleys, New Jersey and New York. Data modified from Witte (1997), Connally and others (1989), Stanford and Harper (1985), and Ridge (1983).

CONCLUSIONS

End moraines in northwestern New Jersey were deposited at the margins of active ice lobes. They represent places where the glacier's terminus remained at a relatively constant position for one thousand to fifteen hundred years for the Terminal Moraine and several hundreds of years for the larger recessional moraines. They largely consist of till, formerly ice-entrained basal debris carried to the glacier's terminus, where it is released by melting. Over time the accumulation of debris and its redistribution across the glacier's periphery chiefly by mass wasting, and to a much lesser extent, ice thrusting, resulted in differential melting and stagnation of the glacier's marginal zone. Moraine-parallel ridges may have formed by ice shove or they are colluvial ramparts formed where debris was shed off the glacier's terminus. The Terminal Moraine and the Ogdensburg-Culvers Gap and Augusta moraines were laid down following a readvance of the Kittatinny Valley lobe. Marginal stagnation was caused by irregular topography and also by burial of the glacier's terminus, and not extensive melting and downwasting.

The reconstruction of glacial-lake histories, and delineation of ice-retreat positions marked by end moraines and outwash heads of ice-contact deltas show that the margin of the Kittatinny Valley lobe retreated in a systematic manner to the northeast. The interpretation of changes in ice flow during deglaciation and the presence of readvances marked by some end moraines shows live ice was present or not very far from the retreating stagnant margin throughout deglaciation. This evidence is in strong contrast to the concept of deglaciation by regional stagnation or valley ice-lobe stagnation as suggested by Connally and others (1989).

Five ice margins, the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex have been identified that mark major recessional positions of the Kittatinny Valley lobe.

The Sparta, Culvers Gap, Augusta, and Sussex margins are traceable onto the New Jersey Highlands, and all margins, except Sparta, are traceable westward into Minisink Valley. All appear to represent halts in ice retreat, and some mark minor readvances of the Kittatinny Valley lobe. This pattern of ice retreat is different from the more rapid style of retreat postulated for the lower part of Kittatinny Valley. These differences as well as the close spacing of the Culvers Gap and Augusta margins, their large traceable extent, and correlation with extensive end moraines and ice-contact deltas in western New Jersey may indicate that other factors, besides local topographic control, have influenced the retreat history of the Kittatinny Valley lobe. Although the O^{18} record suggests that the climate remained relatively cold and stable during the few thousand years that it took for ice to retreat from New Jersey, there exists minor fluctuations that may have influenced deglaciation.

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CHAPTER IV

GLACIAL AQUIFERS OF NEW JERSEY

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ABSTRACT

The principal glacial aquifers of New Jersey are stratified valley-fill deposits greater than 100 feet thick. There are four main types of valley fill. *Unfilled glacial lake basins* contain sand and gravel deposited in lacustrine fans and deltas; and silt, clay, and fine sand deposited on lake bottoms. The lacustrine-fan deposits, where buried by lake-bottom deposits, may be confined aquifers. *Filled glacial lake basins* include these same sediments plus a fluvial sand and gravel at the surface. Both confined and unconfined aquifers may be present. *Fluvial valleys* do not contain significant silt and clay. If sufficiently thick, fluvial valley fills are unconfined aquifers. *Complex valley fills* occur along the late Wisconsinan terminal moraine, where till and lake deposits of two glaciations are stacked. There may be multiple aquifers and confining beds in these valley fills.

INTRODUCTION

Glacial aquifers provide approximately 40 percent of the ground-water supply of northern New Jersey (Hoffman and. Mennel, 1997). They are the sole or primary source for many public-water supplies. Glacial aquifers include some of the most productive aquifers in the state, but their small size makes them susceptible to depletion and their locally high permeability and proximity to the surface make them susceptible to pollution. This chapter describes the geology of these aquifers and how their geology controls the occurrence and movement of ground water.

In New Jersey, glacial aquifers generally have, in the past, been treated as consisting of a single, albeit highly variable, sediment, variously identified as "valley fill" or "stratified drift". This approach is adequate for regional aquifer studies but is insufficient for a detailed understanding of individual valley fills. A variety of permeable and semipermeable glacial materials deposited in glacial streams and lakes and by glacial ice comprise these aquifers and are an important control on the quantity and movement of water in them. Understanding the environments of deposition and sedimentary characteristics provides a rational basis for mapping the aquifers, determining their properties, and managing their ground-water resources.

This chapter will first describe the sedimentary architecture of the valley-fill deposits. It will then review and summarize published hydrologic data for the glacial deposits. Lastly, it will discuss ground-water flow and recharge. This chapter is modified from a text for a Cook College Office of Continuing Professional Education field course on the hydrogeology of glacial deposits in New Jersey (Stanford and Ashley, 1992).

VALLEY-FILL SEDIMENTS

The principal glacial aquifers in New Jersey are stratified valley-fill deposits. Almost all are of late Wisconsinan age but a few are Illinoian. Figure IV-1 shows the location of the principal aquifers and major well fields. Aquifers capable of supplying large-capacity wells are generally restricted to valley fills greater than 100 feet thick. Productive aquifers in valley fills less than 100 feet thick occur in a few places where permeable deposits can receive recharge from surface-water bodies or underlying bedrock units. There are four main types of valley-fill: (1) unfilled glacial lake basins, (2) filled glacial lake basins, (3) fluvial valleys, and (4) complex

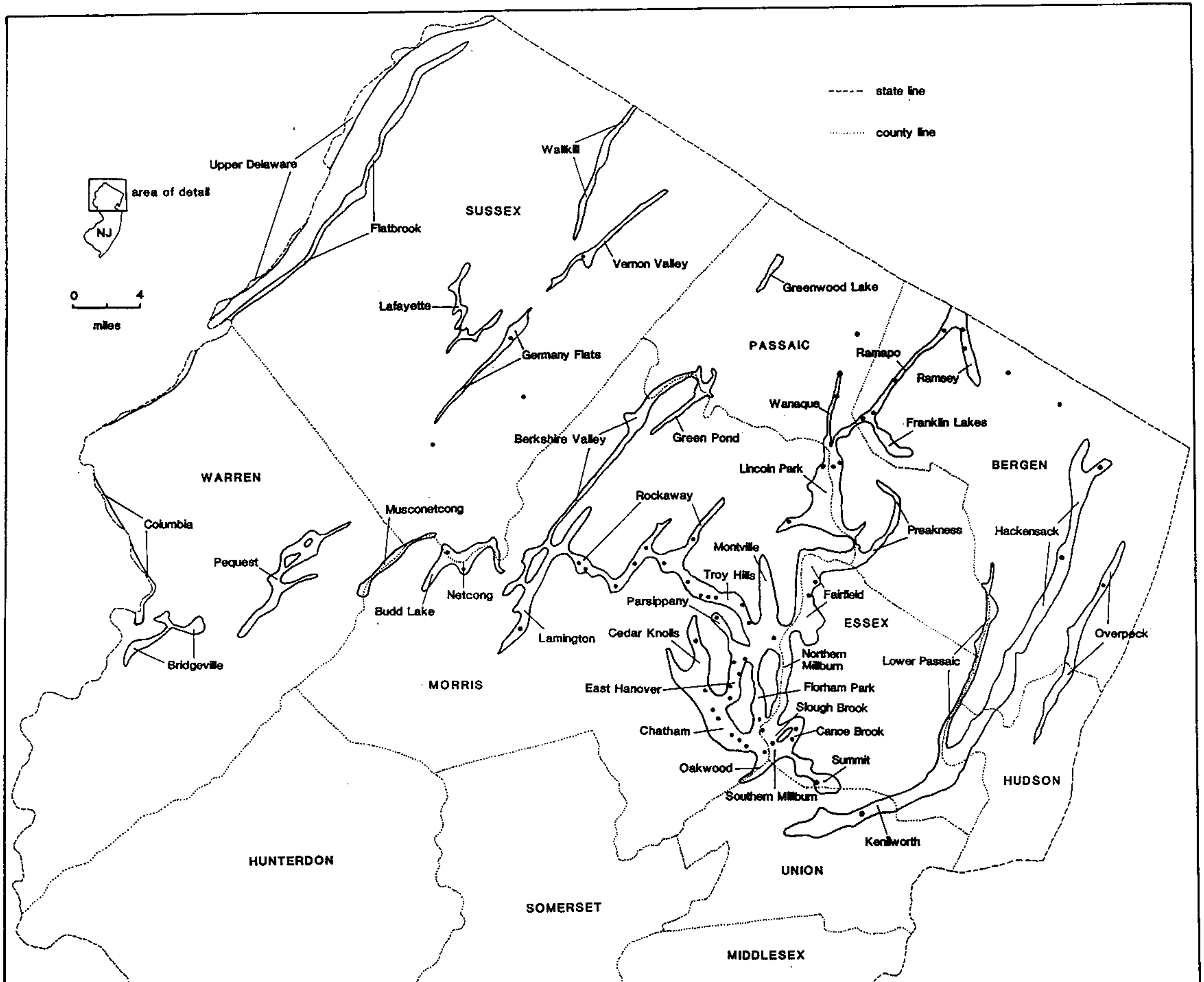


Figure IV-1. Principal valley-fill aquifers of New Jersey. These aquifers occur where permeable valley-fill sediment is generally greater than 100 feet thick. Dots are locations of major wells or well fields drawing from these aquifers. Aquifers in the central Passaic River basin are from Hoffman and Stone (1991); elsewhere, they are drawn based on thickness and sediment information from Stanford and others (1990). Some names are modified from Nichols (1968a), Vecchioli and Miller (1973), Nemickas (1974, 1976), and Canace and others (1993).

aquifer systems along the terminal moraine. These are described in turn below.

Unfilled Glacial Lake Basins

These lake basins were either too large, too deep, or received too little sediment to fill completely with deposits. They include both the large sediment-dammed and large bedrock-dammed lakes described in Chapter II. Most of the glacial lakes in major valleys in New Jersey are in this category. Landforms in these lakes (fig. IV-2) are characterized by deltas and lacustrine fans alternating with low-lying lake plains. The lake plains are today mostly covered by wetlands and floodplains. Figure II-8 in Chapter II provides a surficial geologic map and section of such a setting in a part of glacial Lake Wallkill near Sussex.

The distribution of meltwater deposits in these basins is shown in figure IV-2. Deltas, fed by sediment-laden meltwater discharging from the glacier in subglacial tunnels and in channels along the ice margin on uplands, built out into the lake from stable ice margins during glacial retreat. Sand and gravel were deposited in the deltas as inclined layers below lake level (foreset beds) and as horizontal fluvial layers (topset beds) above lake level. Deltas generally coarsen upward from sand and pebbly sand foreset beds to pebble and cobble gravel topset beds, and their distal parts may overlie and interfinger with silt and fine sand lake-bottom sediment. Deltas may also contain minor beds of poorly sorted debris-flow sediment deposited by mass flow of material from the glacier or from the unstable slopes of the deltas themselves. Between major ice-margin positions, and at the base of deltas at major ice margins, fans of sand and gravel were deposited where subglacial tunnels discharged meltwater into the lake. Bedding and grain size are more variable in the fans than in deltas, and beds of debris-flow sediment are more common. In quiet water away from the ice margin, silt and clay and, in places, fine sand, settled on the lake bottom. This fine sediment gradually filled the lowest parts of the lake and covered much of the fan sediment.

The resulting distribution of permeable sand and gravel bodies and relatively impermeable silt and clay bodies provides the geologic framework for both confined and unconfined aquifers. The lacustrine-fan deposits, where covered by lake-bottom sediment, are confined aquifers in many places. They supply some of the highest-yielding wells in the state. However, in large lake basins they occur only on a small part of the lake bottom, generally as northerly trending linear bodies that track the position of the retreating tunnel mouth. The deltaic deposits and the outcropping parts of lacustrine fans are not covered by lake-bottom sediment and therefore are unconfined aquifers in places where there is a sufficient saturated thickness. Because the deltas and outcropping lacustrine fans may be continuous in the subsurface with the confined lacustrine-fan aquifers, their outcrops may be important recharge areas for those aquifers, particularly where they are in contact with a stream, lake, or wetland. The lake-bottom sediment, because it is relatively impermeable, continuous, and thick, is a confining or leaky confining layer.

Filled Glacial Lake Basins

These lake basins were sufficiently small, shallow, or received enough glacial sediment to fill completely with deposits. Figure IV-3 illustrates the setting, and figure IV-4 provides a map and section for a typical filled basin along the Rockaway River. They include both the sediment-dammed lakes in narrow valleys and the small bedrock-dammed lakes described in Chapter II. Landforms of sediment-dammed lakes (figs. IV-3 and IV-4) are, typically, eroded delta remnants along valley walls separated by fluvial terraces on the valley bottom. Small bedrock-dammed lakes are characterized by deltas that have nearly filled the lake basin. Generally, both types lack exposed lake-bottom plains.

In filled sediment-dammed lake basins (fig. IV-3), as in the unfilled glacial lake basins previously described, deltas build outward into the lakes from ice margins, lacustrine-fan sediment is deposited on the floor of the lake, and lake-bottom sediment fills the low areas. The lake-bottom sediment is generally more sandy than in the unfilled lake basins because these short-lived lakes do not trap silt and clay as effectively as do the large, stable lakes. As the ice margin retreats the lake bottom is gradually exposed as the accumulating lake-bottom and deltaic sediment fills the basin. The lake bottom can also become exposed as the lake level drops when the sediment and ice blocks damming the lake erode or melt. Streams of meltwater flowing across the exposed bottom of the lake lay down fluvial sand and gravel on top of the lake sediment. This completes the three-part layering typical of filled sediment-dammed lakes: an uppermost fluvial sand and gravel overlies lake-bottom and deltaic sand, silt, and minor clay, which in turn generally overlie a basal lacustrine-fan sand and gravel.

This sediment distribution again provides the geologic framework for both confined and unconfined aquifers. The uppermost fluvial sand and gravel is generally unconfined. Where it is in hydrologic connection with streams, lakes, or wetlands it may be an aquifer. The lake-bottom sediment is a leaky confining layer. The basal lacustrine-fan deposits are,

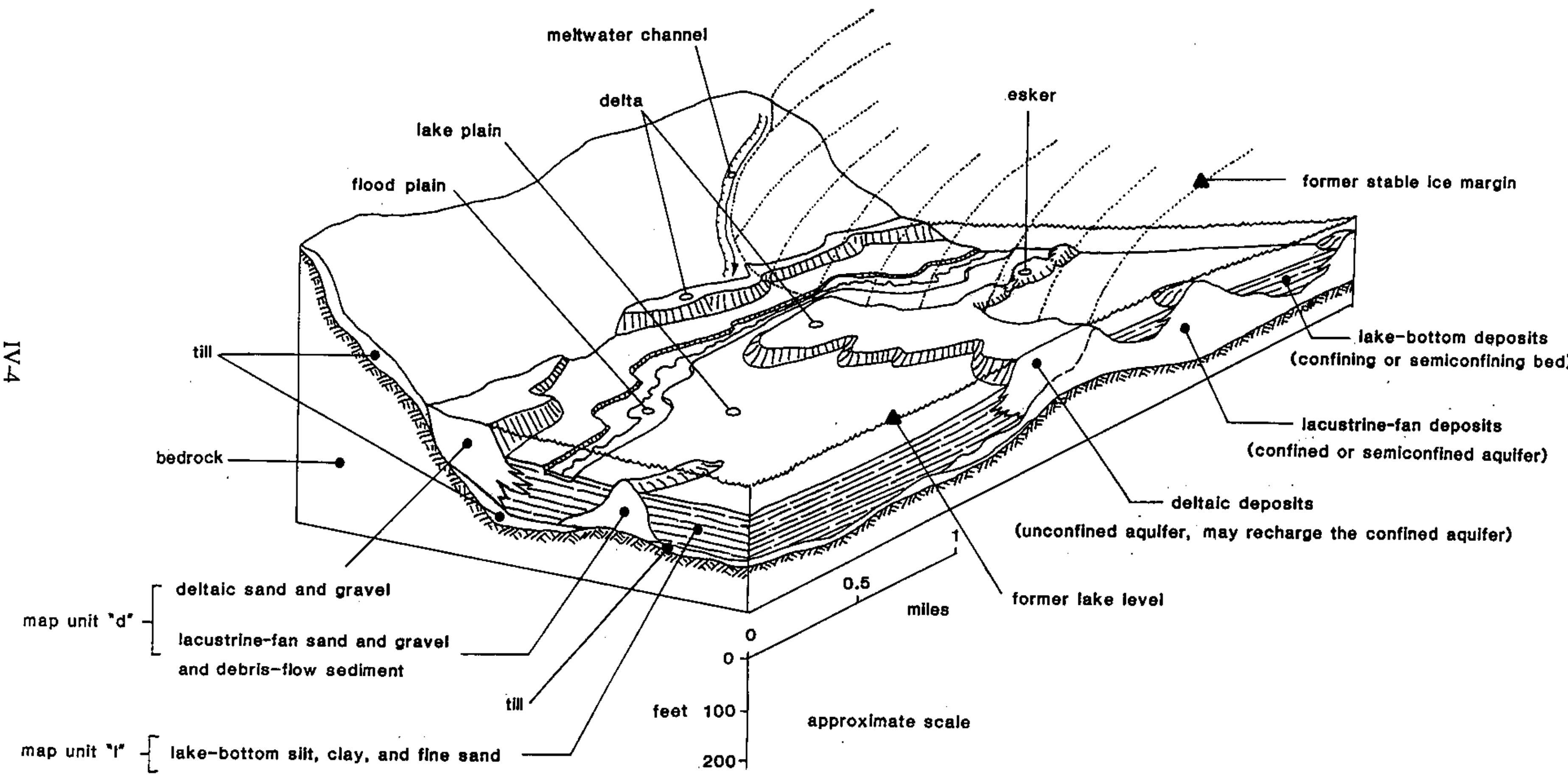


Figure IV-2. Block diagram of an idealized unfilled glacial lake basin. Open circles indicate landforms, filled circles indicate sediment. The hydrologic character of the sediment is indicated at the right. Triangles indicate former ice margins and lake levels during deglaciation.

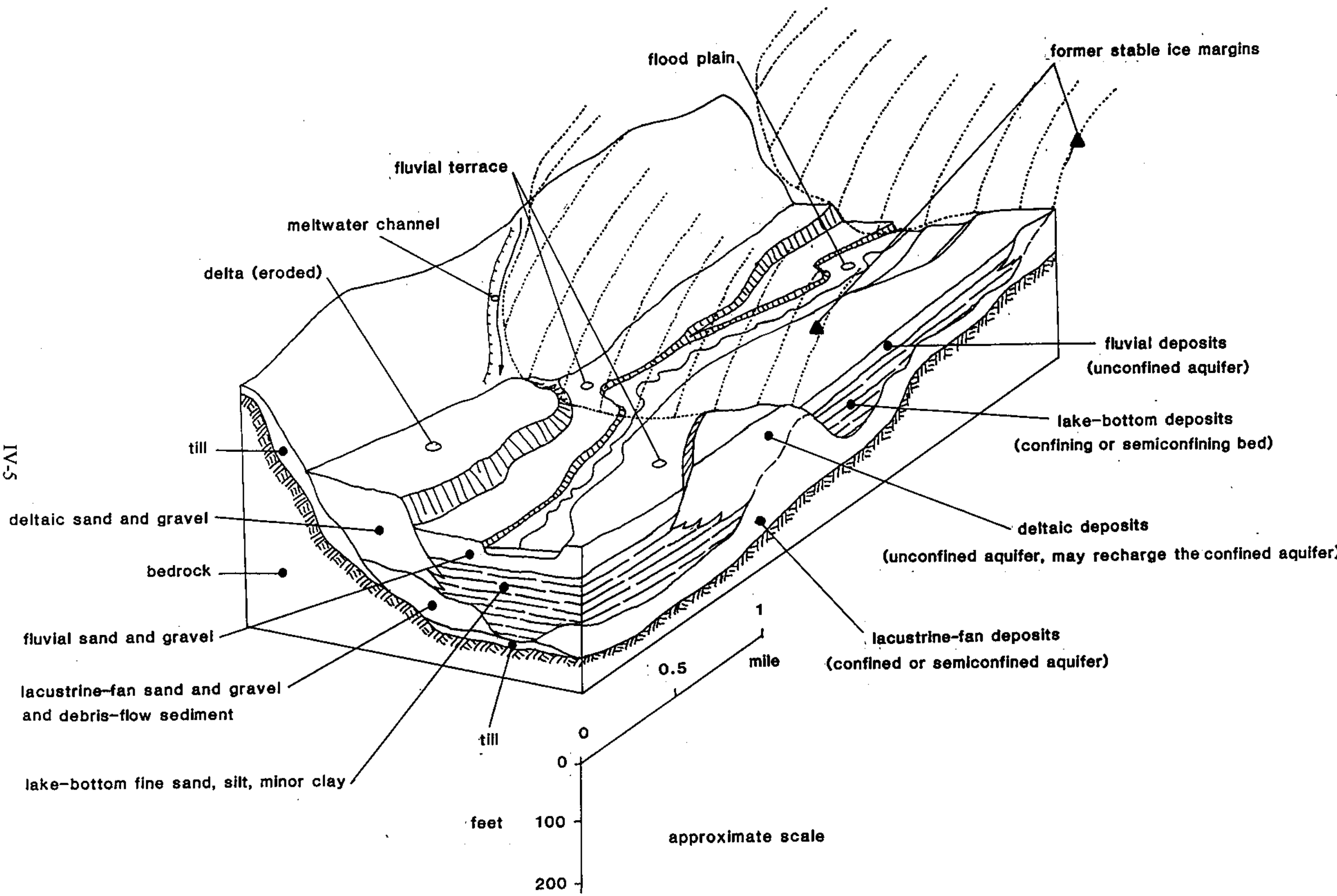


Figure IV-3. Block diagram of an idealized filled, sediment-dammed glacial lake basin.

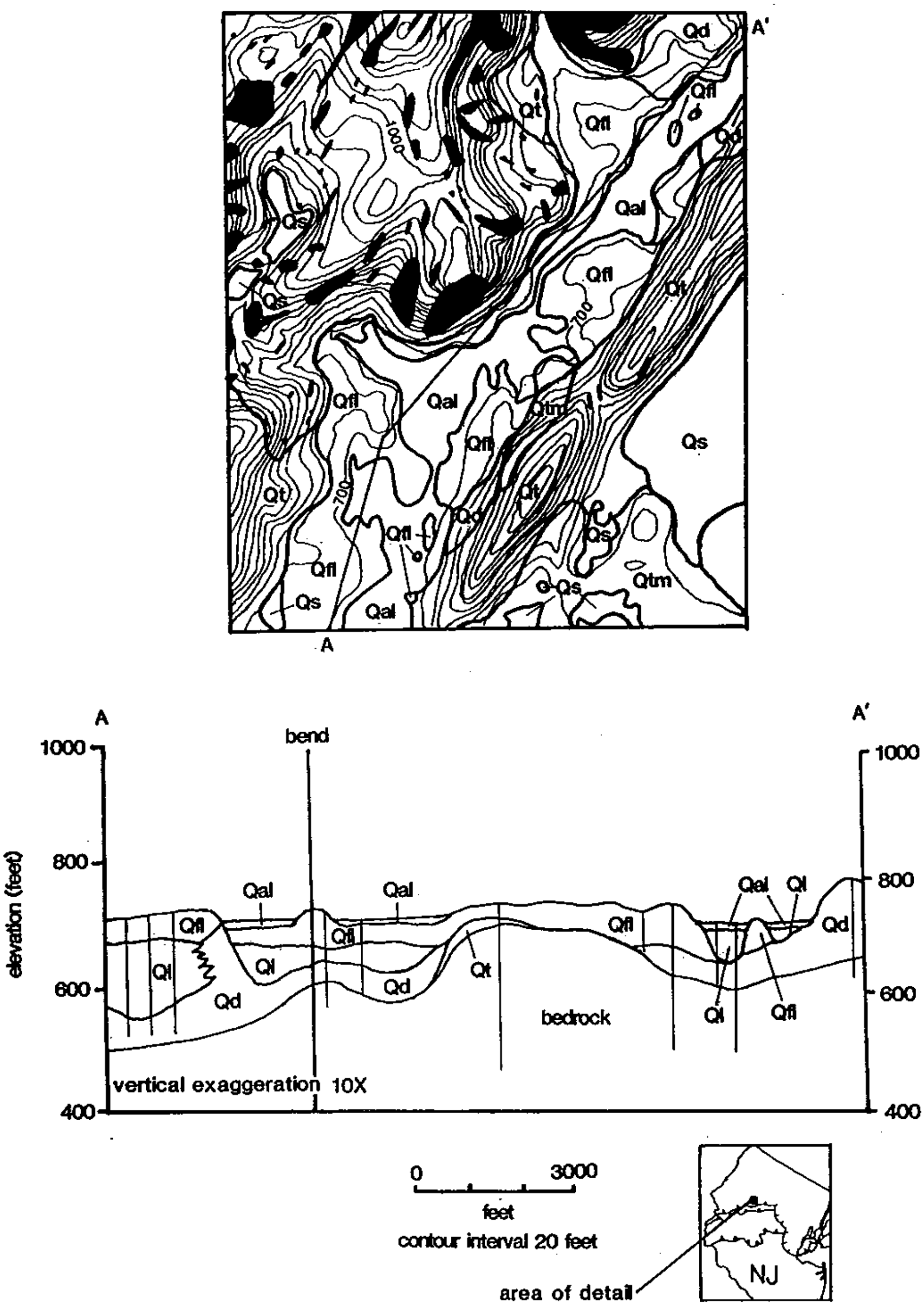


Figure IV-4. Map and section of a filled, sediment-dammed glacial lake basin near Kenvil, New Jersey. Units are: Qt=till, Qtm=terminal moraine deposits, Qd=deltaic and lacustrine-fan sand and gravel, Ql=lake-bottom silt and fine sand, Qfl=fluvial and deltaic sand and gravel, Qs=swamp deposits, Qal=alluvium. Unlabelled areas are thin, discontinuous till. Black areas are bedrock outcrops. Vertical lines on section A-A' mark wells and test borings. Topography from U. S. Geological Survey Dover 7.5 minute quadrangle. Geology from Stanford (1989).

as in the unfilled lake basins, productive confined aquifers in places. The deltas may also be unconfined aquifers and, because they are continuous in the subsurface with the lacustrine-fan deposits, in places may transmit recharge to the confined lacustrine-fan aquifers.

Filled bedrock-dammed lakes generally lack the three-part layering described above. Rather, the fill consists of deltaic foreset sand and a thin topset sand and gravel at the surface. Lake-bottom and lacustrine-fan deposits are generally scant or absent, because there is little trapping of silt and clay and little subglacial drainage in the small upland valleys where these deposits occur. Thus, these deposits are unconfined but are rarely extensive or thick enough to be aquifers.

Fluvial Valleys

Valleys that sloped away from the retreating ice margin did not contain glacial lakes. Instead, meltwater drained down these valleys as streams. Typical landforms are terraces of sand and gravel on the valley bottom, separated by scarps (fig. IV-5). Figure IV-6 is a surficial geologic map and section of such a valley.

Figure IV-5 illustrates the distribution of meltwater deposits in fluvial valleys. Meltwater streams deposited plains of sand and gravel from successive ice margins during glacial retreat. These plains rise in elevation and coarsen in grain size towards "heads of outwash", which mark the former position of the ice margin. Streams depositing plains from later ice-margin positions eroded into earlier plains, forming terraces. The deposits in each plain generally coarsen from sand and pebbly sand far from the ice margin to cobble gravel near the ice margin. Poorly sorted debris-flow sediment may be interbedded with the gravel at the ice margin. In places on the valley bottom where the glacier had scoured basins in the bedrock, small deposits of deltaic and lake-bottom sand and silt may underlie the fluvial sand and gravel.

Silt and clay generally remain in suspension in the streams and are flushed from the valleys except in the rare valley-bottom basin. Thus, fluvial valley fills consist almost entirely of sand and gravel and therefore form unconfined aquifers if sufficiently thick. Sediment in fluvial valleys is generally not as thick as it is in valleys that contained glacial lakes, and in most cases productive aquifers occur only where the sediment is in hydraulic connection with a surface-water body.

Complex Aquifers Along the Terminal Moraine

In some valleys and lowlands along the terminal moraine between Millburn and Belvidere, late Wisconsinan ice overlapped but did not erode previously deposited Illinoian valley-fill sediment. Glaciofluvial and glaciolacustrine deposits and till of late Wisconsinan age were deposited on top of similar deposits of Illinoian age, forming a complex valley fill. Figure IV-7 illustrates this setting, and figure IV-8 is a map and section of such an area between Denville and Parsippany.

In the upper Passaic basin, Rockaway valley, Lamington valley, and, possibly, the Pequest valley, glacial lakes formed during both the Illinoian and late Wisconsinan glaciations. Upon retreat of Illinoian ice, deltaic, lacustrine-fan, and lake-bottom deposits partially filled these valleys. As late Wisconsinan ice advanced into this landscape, lakes again flooded the valleys. In a narrow zone at its farthest advance, late Wisconsinan ice ramped up over the Illinoian deposits, which were completely eroded by the glacier in valleys farther north. Till and lacustrine deposits were stacked on top of Illinoian sediment. As lakes drained, fluvial sediment was deposited on the lacustrine deposits in places.

The resulting valley fill forms the framework for both confined and unconfined aquifers. Permeable lacustrine-fan, deltaic, and, in places, fluvial deposits covered by relatively impermeable lake-bottom deposits and till form confined aquifers. Unconfined lacustrine-fan, deltaic, and fluvial sediment at the surface, deposited during final recession of late Wisconsinan ice, are potential aquifers.

Recharge routes are not as straightforward as in the previously described valley-fill environments. Deltas south of the terminal moraine may be continuous in the subsurface with the buried confined aquifers beneath and north of the moraine, and therefore may recharge those aquifers. Some recharge may also occur through sandy till and debris-flow sediment of the moraine, and through the delta, fan, and fluvial deposits north of the moraine where they are in contact with the older permeable deposits in the subsurface.

Along the terminal moraine east of Millburn, in the Musconetcong and Delaware valleys, and in parts of the Pequest valley, there is no significant Illinoian sediment in the subsurface and the complex stratigraphy described above does not occur. Instead, till of the moraine overlies and interfingers with late Wisconsinan fluvial and deltaic deposits, particularly at the outer edge of the moraine, and extends beneath lacustrine sediment in valleys back from the inner edge of the moraine. This stratigraphy also occurs on

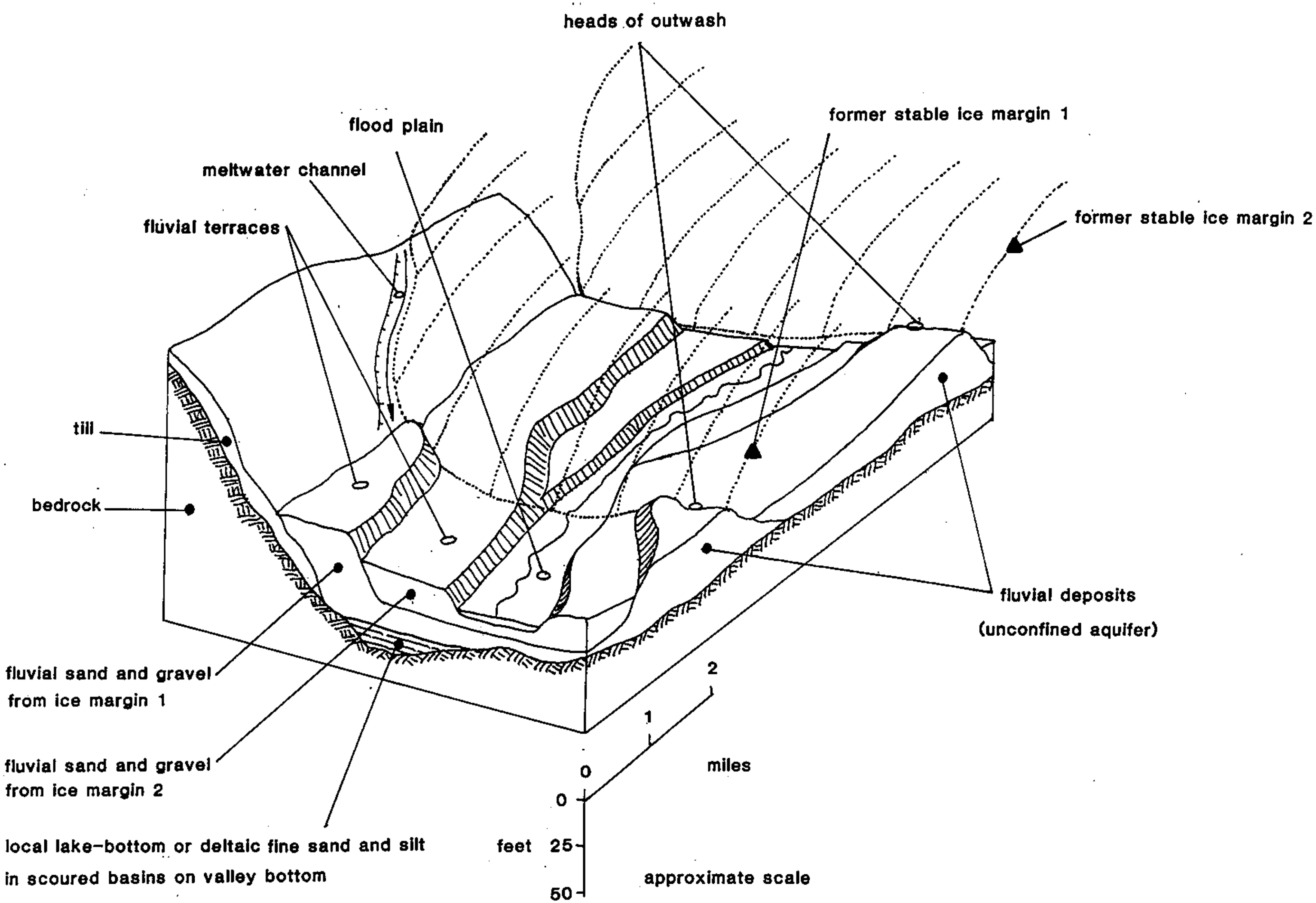
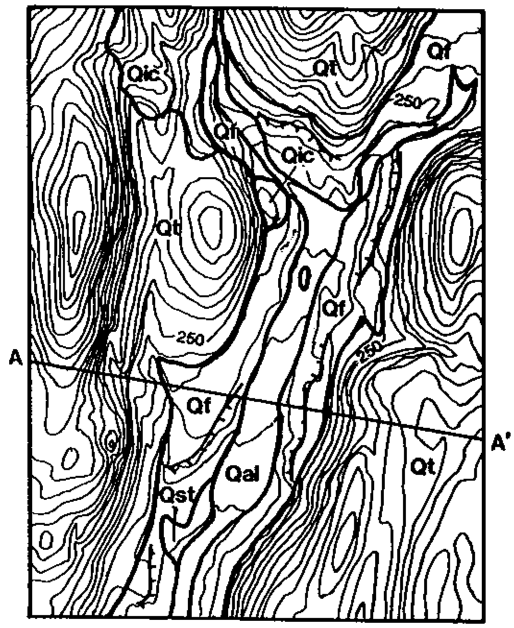


Figure IV-5. Block diagram of an idealized fluvial valley.



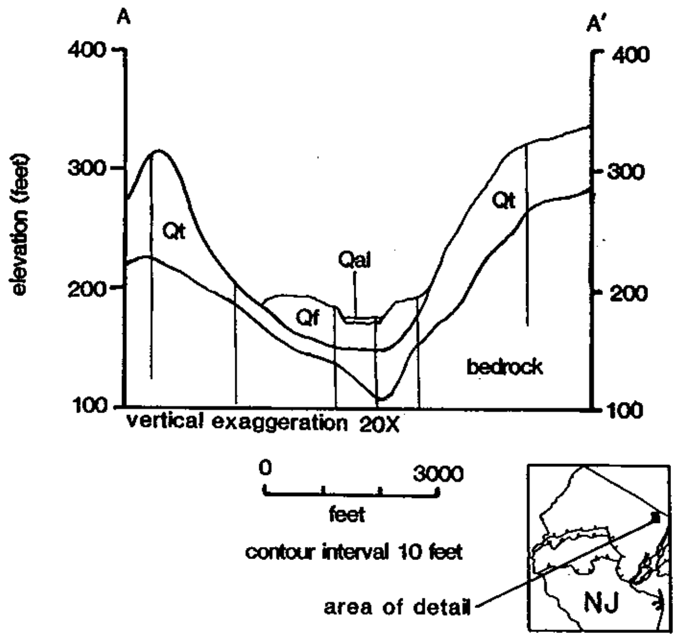


Figure IV-6. Map and section of a typical fluvial valley near Upper Saddle River, New Jersey. Units are: Qt=till, Qf=fluvial sand and gravel, Qic=ice-contact deposits, Qal=alluvium, Qst=stream terrace deposits. Hachured lines are fluvial scarps. Topography from U. S. Geological Survey Park Ridge 7.5 minute quadrangle.

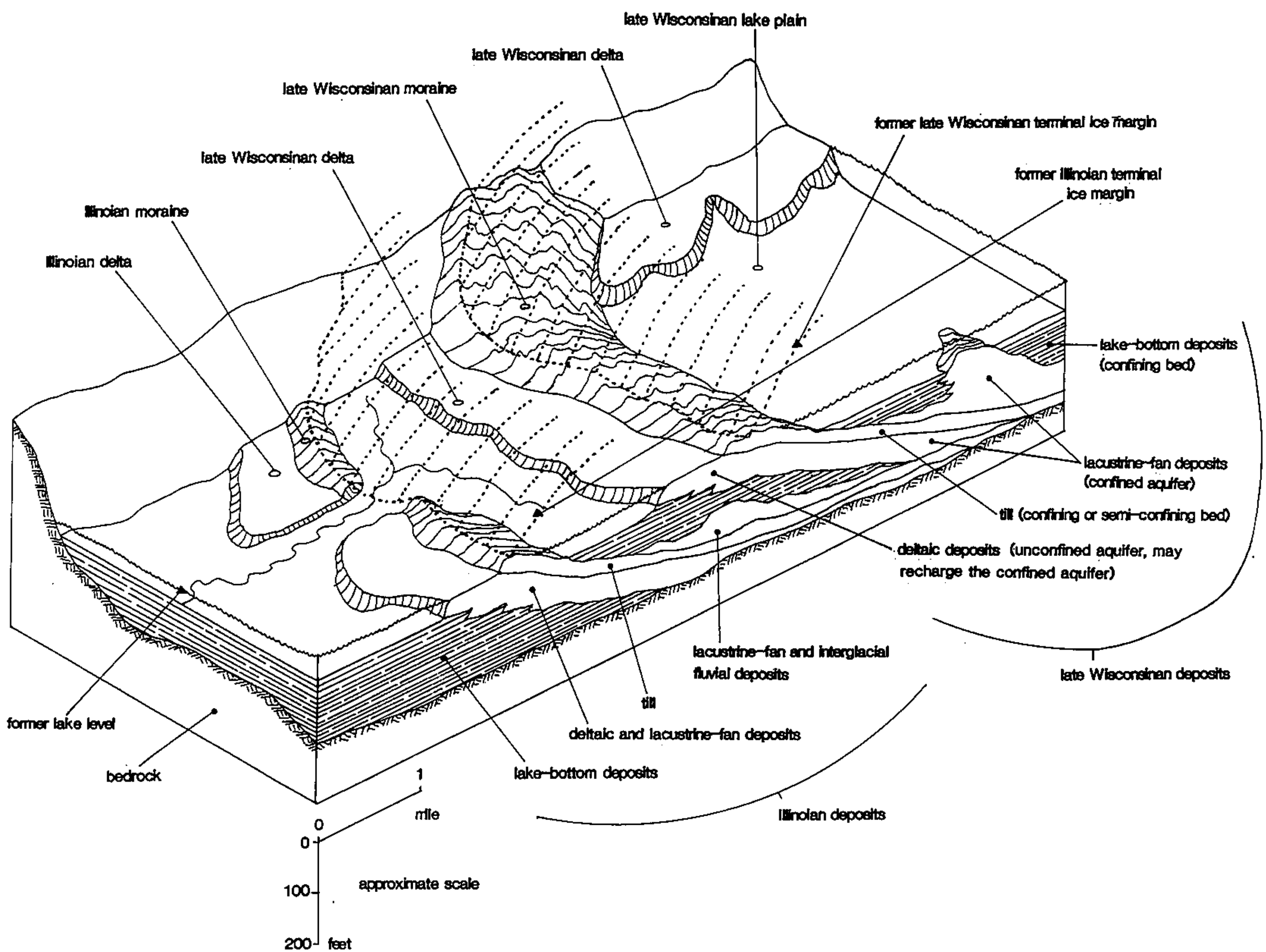


Figure IV-7. Block diagram of an idealized complex valley fill along the terminal moraine.

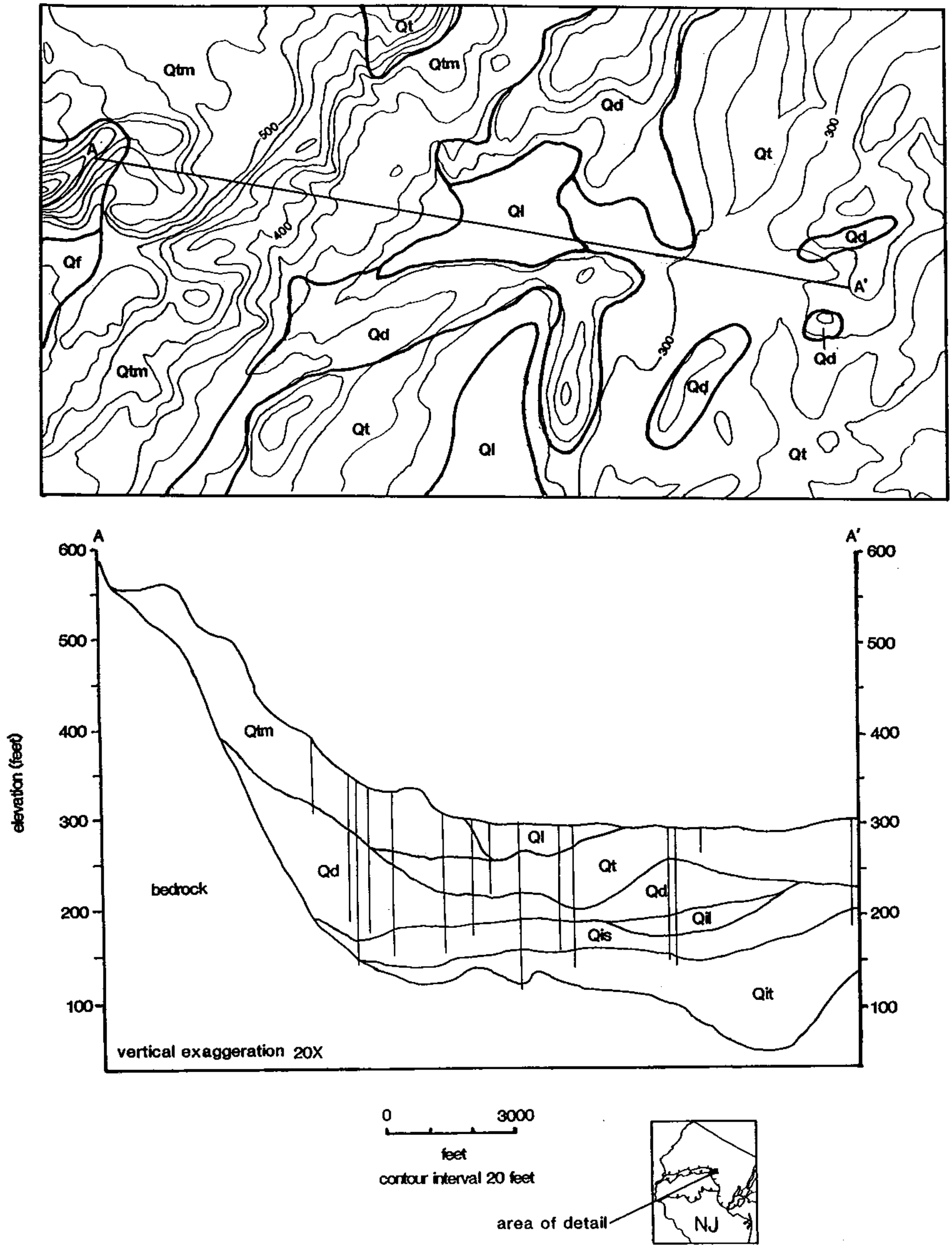


Figure IV-8. Map and section of a complex valley-fill along the terminal moraine. Illinoian units are: Qit=till, Qis=deltaic and lacustrine-fan sand and gravel, Qil=lake-bottom silt and clay. Late Wisconsinan units are: Qd=deltaic and lacustrine-fan sand and gravel, Qt=till, Qtm=terminal moraine deposits, Ql=lake-bottom silt and clay, Qf=fluvial sand and gravel. Unlabelled area is weathered gneiss bedrock. Topography from U. S. Geological Survey Morristown and Boonton 7.5 minute quadrangles. Geology from Canace and others (1993).

a smaller scale at some of the recessional moraines north of the terminal moraine.

Other Glacial Aquifers

Very few water wells tap glacial sediment other than in the valley-fill settings described above. A few domestic wells are reported to draw water from till in drumlins, although some may draw from granular weathered bedrock or stratified sand and gravel in the cores of the drumlins. A few other wells are completed in bouldery, sandy, poorly-stratified, ice-contact sediment. These deposits form along recessional ice margins, or in areas of stagnant ice, and may include debris-flow sediment from melting ice and collapsed, deformed lacustrine sediment. They occur at only a few places in the state (unit "ic" on fig. II-3). They generally form well-drained ridges that are too dry to supply water. In a few of the deeper valleys in the Highlands, however, these deposits occur in valley bottoms and may be local unconfined aquifers.

HYDROLOGIC PROPERTIES

The previous section outlines the various types of glacial valley-fill aquifer in New Jersey. Each valley has a unique geography and a unique history of glacial lake levels, meltwater-stream drainage, ice-margin positions, and glacial sediment sources. For this reason, the exact configuration of water-producing beds, confining beds, and principal recharge routes varies from valley to valley. Thus, a thorough understanding of the hydrology of glacial valley-fill aquifers requires detailed geologic and hydrologic study of individual valleys. The purpose of this section is to summarize the results of published hydrogeological investigations in New Jersey in order to provide a range of yields and specific capacities for wells tapping glacial aquifers, and of hydraulic conductivities, storativities, and ground-water flow paths for the aquifers. The well-construction, aquifer-test, and water-level data on which these summaries are based are available in the publications listed in the figure captions and tables. These data are from a variety of sources and their accuracy and completeness are variable. Additional unpublished data, not included in these summaries, are on file at the New Jersey Geological Survey.

Well Yields

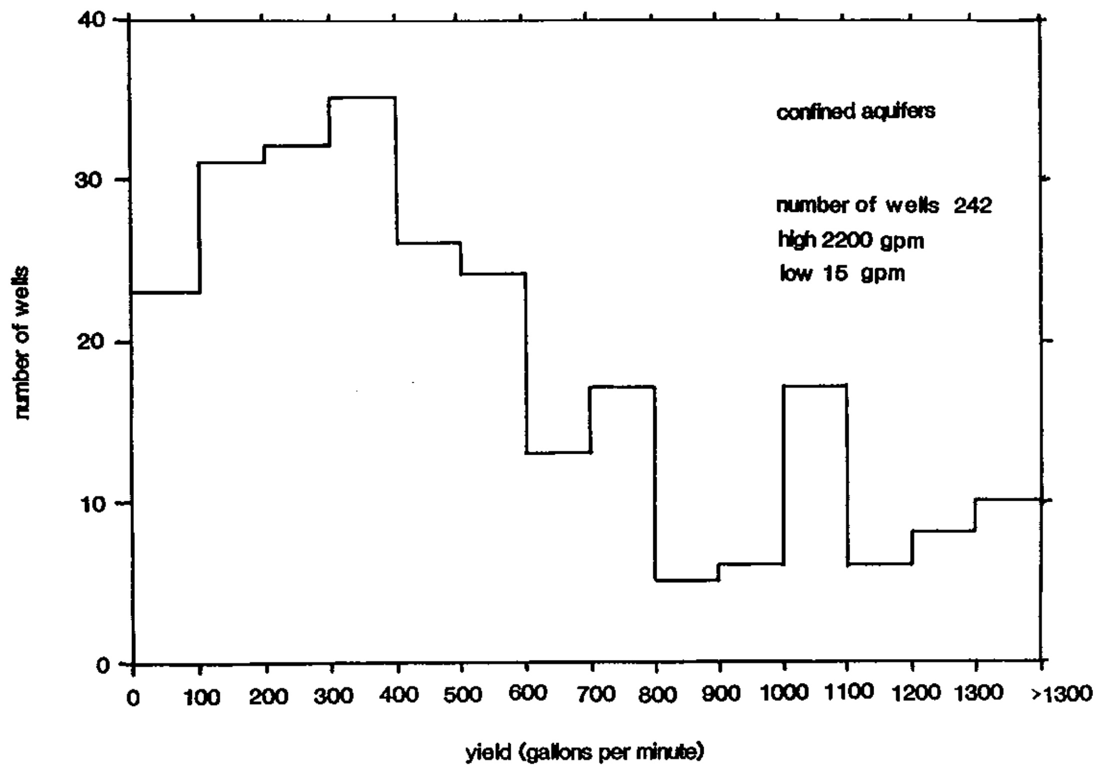
The yield of a well is the volume of water that can be pumped from it in a given unit of time without pumping it dry. It depends largely on the water-bearing properties of the aquifer and on the construction characteristics of the well. For this reason, well yields are commonly reported in two categories: those of domestic wells, which are designed to provide small quantities of water and generally have diameters of 6 inches or less, and those of large-capacity wells, which are designed to provide large quantities. These include industrial and public-supply wells and generally have diameters of 8 inches or more.

Figure IV-9 shows the distribution of yields of large-capacity wells in both confined and unconfined glacial aquifers. Yields of 242 such wells completed in confined aquifers range from 15 to 2200 gallons per minute (gpm). Most of these wells yield between 50 and 550 gpm. Yields for 75 large-capacity wells completed in unconfined aquifers range from 6 to 1750 gpm. Most of these wells yield between 50 and 250 gpm. More high-yield wells are in confined aquifers, due, in part, to the generally greater transmissivity of the lacustrine-fan deposits that constitute confined aquifers and to the fact that water-level declines decrease the transmissivity of an unconfined aquifer but not that of a confined aquifer. It is also due, in part, to the greater number of large-diameter public supply wells drilled in confined aquifers, and to state regulations which do not allow wells to withdraw water within 50 feet of the land surface in a sand and gravel aquifer. This restriction excludes or limits the use of many unconfined aquifers as a water source.

Specific Capacities

The specific capacity of a well is defined as its yield divided by the drawdown in water level in the well caused by the pumping. It is a measure of the ability of the aquifer to transmit water to the well, although, like the yield, it, too, is affected by the well-construction characteristics. Large-capacity wells generally have a larger specific capacity than domestic wells in the same aquifer because their construction characteristics provide greater efficiency.

Figure IV-10 shows the distribution of published specific capacities for large-capacity wells. Specific capacities of 202 such wells completed in confined aquifers range from 0.19 to 500 gallons per minute per foot of drawdown (gpm/ft). Specific



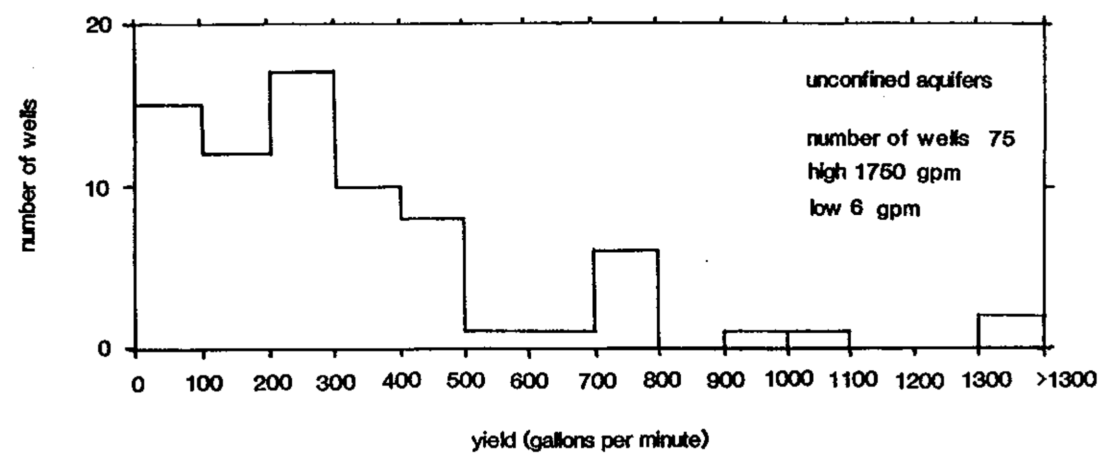
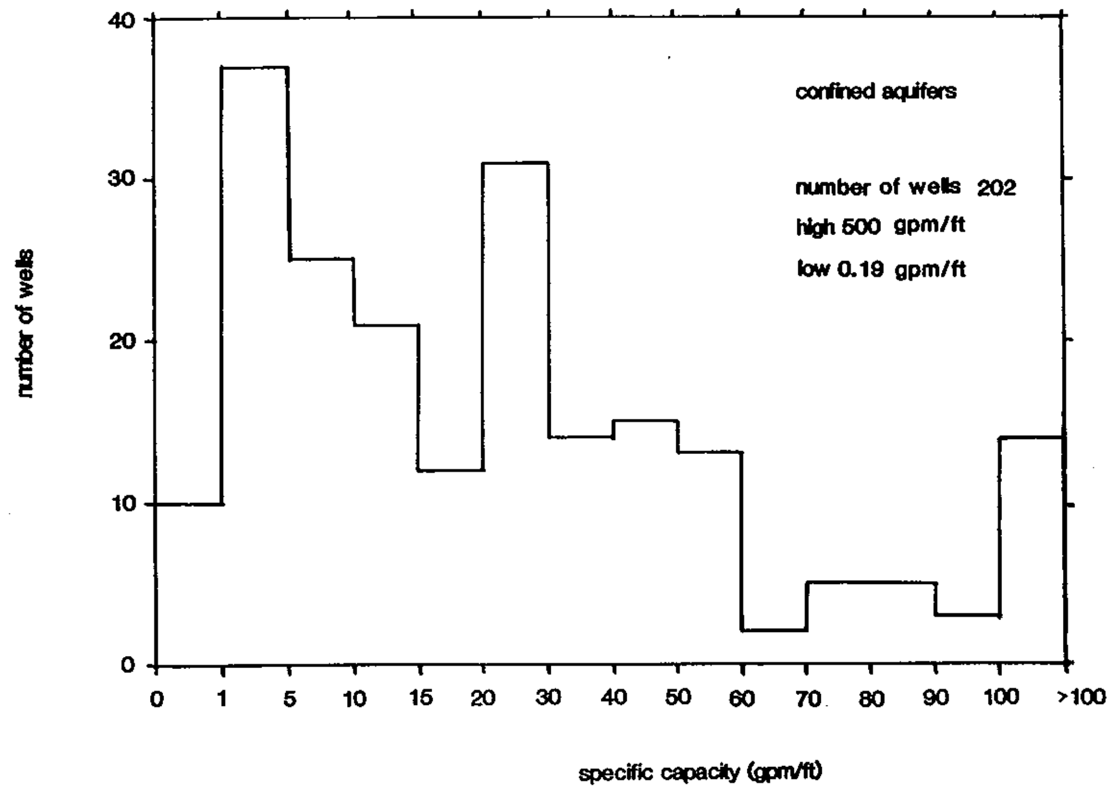


Figure IV-9. Distribution of yields of large-capacity wells completed in glacial sediment in New Jersey. Sources: Herpers and Barksdale (1951), Gill and Vecchioli (1965), Vecchioli and Nichols (1966), Widmer and others (1966), Vecchioli and others (1967), Anderson (1968), Nichols (1968b), Vecchioli and Miller (1973), Miller (1974), Carswell (1976), Carswell and Rooney (1976), Nemickas (1976).



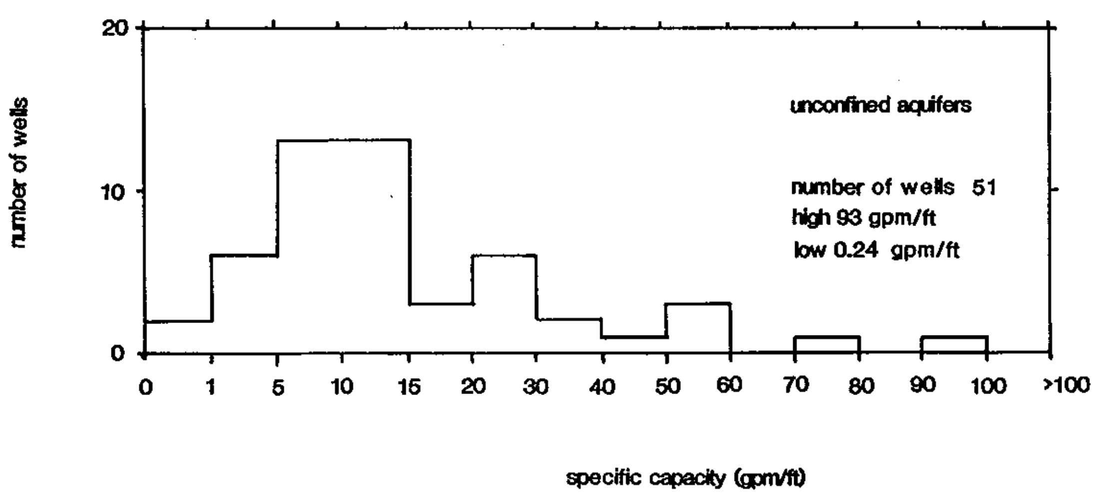


Figure IV-10. Distribution of specific capacities of large-capacity wells completed in glacial sediment in New Jersey. Sources: Herpers and Barksdale (1951), Gill and Vecchioli (1965), Vecchioli and Nichols (1966), Vecchioli and others (1967), Anderson (1968), Nichols (1968b), Vecchioli and Miller (1973), Miller (1974), Carswell (1976), Carswell and Rooney (1976), Nemickas (1976), Canace and others (1983), Hill (1985).

capacities of 51 such wells completed in unconfined aquifers range from 0.24 to 93 gpm/ft. Most of the specific capacities in both groups are between 1 and 30 gpm/ft. As with yields, the greater number of high-specific-capacity wells in confined aquifers is due, in part, to the greater transmissivity of confined aquifers and, in part, to the greater number of deep, large-diameter wells drilled in those aquifers.

Hydraulic Conductivity and Transmissivity

The hydraulic conductivity is a measure of the ability of an aquifer to transmit water. Specifically, it is defined by Darcy's Law, which is an empirically-derived expression for flow of water through porous media. The most common version of the law states: Q=KIA where Q=the discharge rate of water (volume per unit time) flowing through a medium with cross-sectional area A in response to a hydraulic gradient I. K is the hydraulic conductivity of the medium and is in units of length over time. For confined aquifers the transmissivity (T), which is equal to the hydraulic conductivity multiplied by the thickness of the aquifer, is commonly used rather than hydraulic conductivity if the thickness of the aquifer is not known.

Hydraulic conductivity is an intrinsic property of the aquifer. An aquifer's horizontal hydraulic conductivity can be determined by analyzing data from aquifer tests, although in many instances the analytic techniques assume conditions that cannot be duplicated in the field. Nevertheless, such tests are the most reliable means of evaluating aquifers and providing the data needed for constructing and calibrating computer models of the aquifers.

Some aquifer tests have been performed on the glacial aquifers of New Jersey, most as part of ground-water-supply studies. Almost all of these tests were conducted on wells finished in deltaic and lacustrine-fan deposits of sand and gravel. Available values for horizontal hydraulic conductivities computed from these tests are summarized in figure IV-11 and table IV-1.

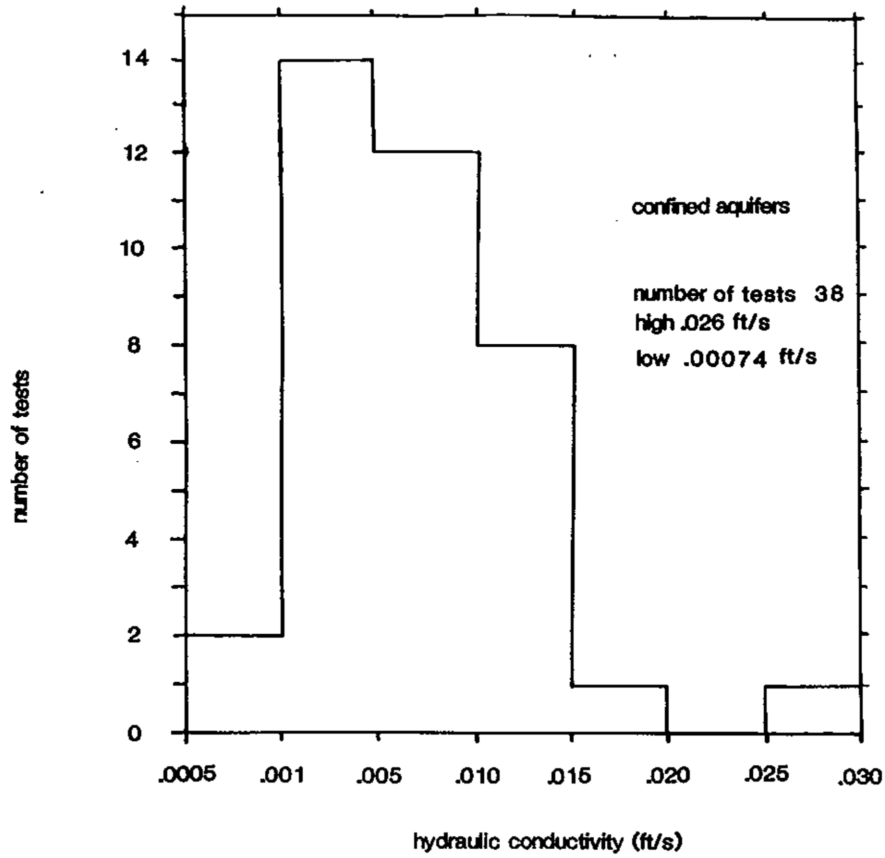
Conductivities of confined aquifers range from 0.00074 to 0.026 foot per second (ft/s); those of unconfined aquifers range from 0.0014 to 0.075 ft/s. Almost all the conductivities in both groups are between 0.001 and 0.015 ft/s. This range is typical of sand and gravel. The hydraulic conductivity of lake-bottom silt and clay confining material has been measured at only a few places in New Jersey (table IV-1). Hill and others (1992) measured a vertical hydraulic conductivity of 3.4×10^{-10} ft/s for silty lake-bottom sediment in the Ramapo valley (fig. IV-1) and Vecchioli and others (1962) reported an average vertical hydraulic conductivity of 3.48×10^{-9} ft/s for lake-bottom silt and clay in the Lake Passaic basin. Meisler (1976) used confining layer vertical hydraulic conductivities of 7×10^{-8} to 4.9×10^{-7} ft/s to calibrate a computer model of aquifers in a part of the Lake Passaic basin. These values indicate that silt and clay is 4 to 8 orders of magnitude less conductive than sand and gravel.

Hydraulic conductivities of till in New Jersey have not been determined by aquifer testing. A literature review by Jeffrey L. Hoffman of the New Jersey Geological Survey (written communication, 1990) indicates that reported hydraulic conductivities for till elsewhere in North America range from 10^{-12} ft/s for compact, non-jointed clay till to 10^{-5} ft/s for sandy till. In the silty-to-sandy till of New Jersey, hydraulic conductivities in the range of 10^{-8} to 10^{-5} ft/s may be expected. Till in New Jersey, because it typically contains a significant amount of sand, generally is not strongly cohesive and so is nonjointed to weakly jointed. Thus, jointing should not be expected to greatly increase hydraulic conductivity.

Storativity

The storativity of an aquifer is the volume of water released by a unit area of the aquifer when subjected to a unit decline in water level. For confined aquifers, storativity is expressed as the "specific storage", which is defined as the volume of water released from a unit volume of the aquifer when subjected to a unit decline in water level. For an unconfined aquifer the equivalent measure is the "specific yield", which is defined as the volume of water released from a unit area of the aquifer when subjected to a unit decline in water level. Like hydraulic conductivity, storativity is an intrinsic property of an aquifer and can be computed by analyzing aquifer-test data. Confined aquifers have much lower storage values (0.0005 to 0.005) than unconfined aquifers (0.1-0.3) because a decline in water level in a confined aquifer reflects mostly release of water due to a decrease in fluid pressure rather than from dewatering. Thus, a given drawdown in a confined aquifer produces less water than the same drawdown in an unconfined aquifer.

Figure IV-12 and table IV-2 show the distribution of available storativity values computed from aquifer tests on wells completed in glacial sediment in New Jersey. As with the hydraulic conductivities, almost all of these wells tap deltaic and lacustrine-fan sand and gravel. As figure IV-12 shows, most of the aquifers are confined or semiconfined. Storativities on the order of 10^{-4} probably represent confined aquifers overlain by thick, continuous silt and clay lake-bottom deposits. Values



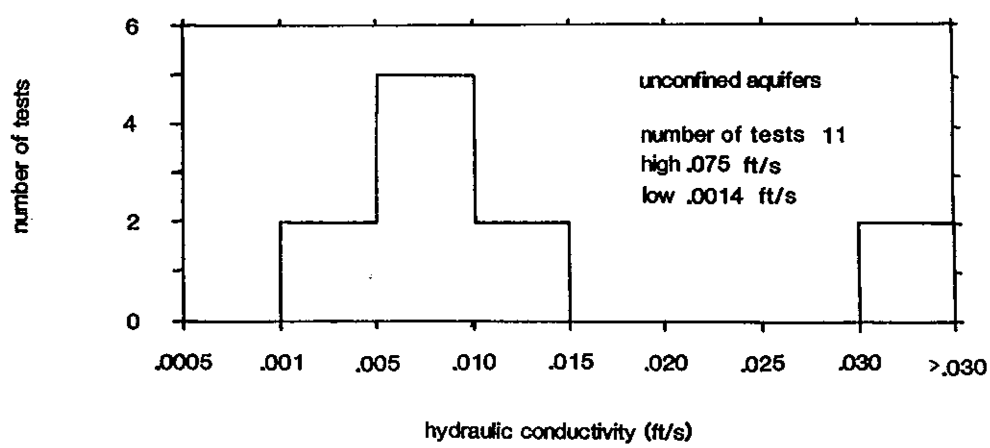


Figure IV-11. Distribution of hydraulic conductivities of glacial sediment computed from aquifer tests. Sources: Gill and Vecchioli (1965), Anderson (1968), Vecchioli and Miller (1973), Meisler (1976), Hutchinson (1981), Canace and others (1983), Hill (1985), Sirois (1986), Hill and others (1992).

Table IV-1. Summary of published hydraulic conductivity data for glacial aquifers in New Jersey.

Hydraulic Conductivity[1] (ft/s)	Type of Data	Location[2]	Reference
CONFINED AQUIFERS			
0.0024	Aquifer test	Rockaway aquifer	Sirois (1986)
0.0039	"	"	"
0.0128	"	"	Gill and Vecchioli (1965)
0.0131	"	"	"
0.0075	"	Ramapo aquifer	Hill and others (1992)
0.0069	"	"	"
0.0045	Laboratory test	"	Vecchioli and Miller (1973)
0.0453	"	"	"
0.003 to 0.004	Model results	Southwestern Essex and southeastern Morris counties	Meisler (1976)
0.0030	Aquifer test	Berkshire Valley aquifer	Canace and others (1983)
0.0110	"	"	"
0.026	Calculated from specific capacity data	Lamington aquifer	Hill (1985)
0.014	"	"	"
0.011	"	"	"
0.016	"	"	"
0.0027	"	"	"
0.0081	"	"	"
0.0091	"	"	"
0.0029	"	"	"
0.0042	"	"	"
0.0083	"	"	"
0.0084	"	"	"
0.016	Aquifer test	"	"
0.0020	"	"	"
0.0068	"	Canoe Brook aquifer	Gill and Vecchioli (1965)
0.0056	"	"	"
0.0061	"	"	"
0.0092	"	"	"
0.0116	"	"	"
0.0115	"	"	"
0.003	"	"	"
0.0018	"	Chatham aquifer	"
0.0065	"	"	"
0.004	"	"	"
0.0033	"	"	"
0.003	"	"	"
0.00074	"	"	"
0.00084	"	"	"
UNCONFINED AQUIFERS			
0.0055	Aquifer test	Pequest aquifer	Hutchinson (1981)
0.0059	"	"	"
0.0114	"	"	"
0.0091	"	"	"
0.0135	"	"	"
0.0014	"	Berkshire Valley aquifer	Canace and others (1983)
0.075	Calculated from specific capacity data	Lamington aquifer	Hill (1985)
0.036	"	"	"
0.0035	Laboratory test	Southeast Union County	Anderson (1968)
0.0082	Aquifer test	Eastern Morris County	Gill and Vecchioli (1965)
0.0077	"	"	"
CONFINING BEDS			
3.4×10^{-10}	Aquifer test	Ramapo aquifer	Hill and others (1992)
3.48×10^{-9}	Laboratory test	Southeastern Morris County	Vecchioli and others (1962)
7×10^{-8} to 4.9×10^{-7}	Model results	Southwestern Essex and southeastern Morris counties	Meisler (1976)
4.7×10^{-8}	Laboratory test	Lamington aquifer	Nicholson and others (1996)

[1]Some hydraulic conductivity values are derived from reported transmissivity values by dividing the transmissivity by aquifer thickness, as determined from geologic logs. Hydraulic conductivities are assumed to be horizontal in aquifers and vertical in confining beds.
[2] Shown on figure IV-1.

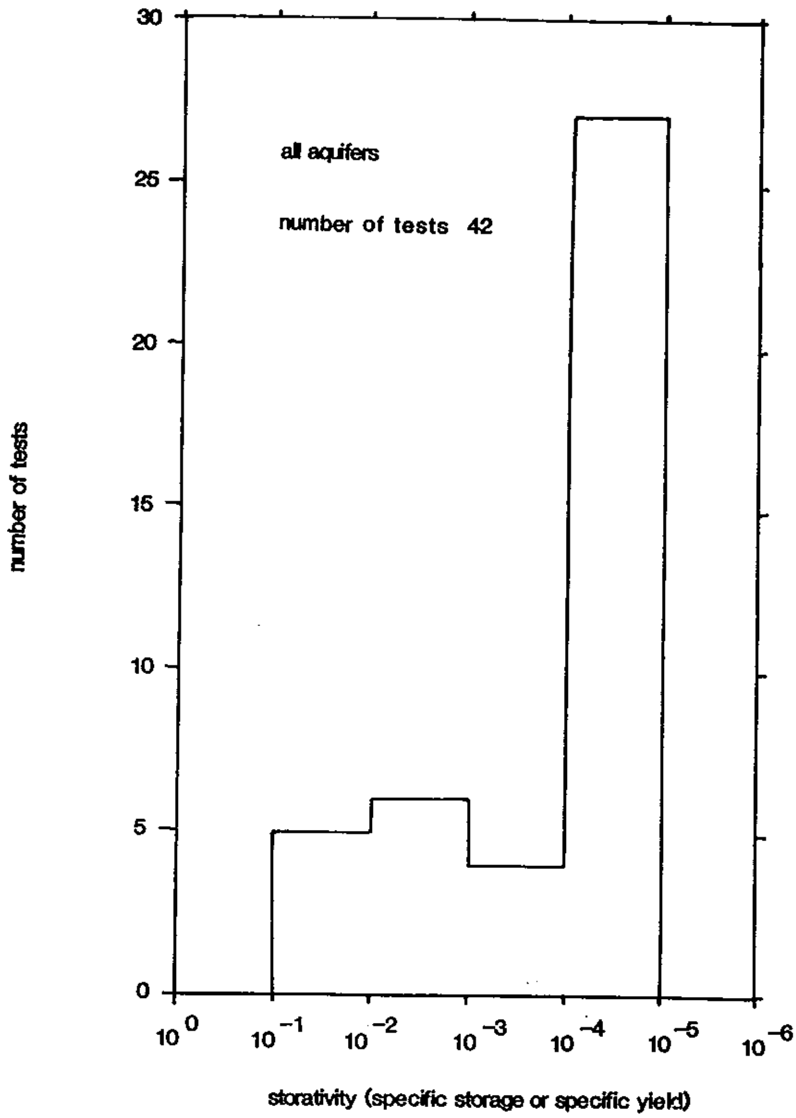


Figure IV-12. Distribution of storativities computed from aquifer tests on wells completed in glacial sediment in New Jersey. Sources as listed for figure IV-11.

Table IV-2. Summary of storativity data for glacial aquifers in New Jersey.

Storativity (dimensionless)	Type of Data	Location[1]	Reference
CONFINED AQUIFERS			
$1.4x10^{-4}$	Aquifer test	Ramapo aquifer	Hill and others (1992)
$1.1x10^{-4}$	"	"	"
$1.0x10^{-4}$	Model results	Southwest Essex and southeast Morris counties	Meisler (1976)
$1.72x10^{-3}$	Aquifer test	Berkshire Valley aquifer	Canace and others (1983)
$2.66x10^{-3}$	"	"	"
$4.9x10^{-3}$	"	"	"
$5.2x10^{-4}$	"	Lamington aquifer	Hill (1985)
$5.0x10^{-4}$	"	"	"
$3.3x10^{-4}$	"	Canoe Brook aquifer	Gill and Vecchioli (1965)
$3.5x10^{-4}$	"	"	"
$4.6x10^{-4}$	"	"	"
$5.8x10^{-4}$	"	"	"
$1.1x10^{-3}$	"	"	"
$2.3x10^{-4}$	"	"	"
$1.5x10^{-4}$	"	Chatham aquifer	"
$3.5x10^{-4}$	"	"	"
$3.3x10^{-4}$	"	"	"
$2.3x10^{-4}$	"	"	"
$2.7x10^{-4}$	"	"	"
$3.8x10^{-4}$	"	"	"
$2.0x10^{-4}$	"	"	"
$2.0x10^{-4}$	"	"	"
$2.0x10^{-4}$	"	"	"
$4.7x10^{-4}$	"	"	"
$5.7x10^{-4}$	"	"	"
$4.0x10^{-4}$	"	"	"
$9.3x10^{-4}$	"	East Hanover aquifer	"
$9.0x10^{-4}$	"	"	"
$5.2x10^{-4}$	"	Rockaway aquifer	"
$5.0x10^{-4}$	"	"	"
UNCONFINED AQUIFERS			
0.03	Aquifer test	Pequest aquifer	Hutchinson (1981)
0.03	"	"	"
0.05	"	"	"
0.02	"	"	"
0.16	Model results	Southwest Essex and southeast Morris counties	Meisler (1976)
0.081	Aquifer test	Berkshire Valley aquifer	Hill (1985)
0.086	"	"	"
0.1 to 0.2	Extrapolation from laboratory tests	Southeast Union County	Anderson (1968)
0.15	Aquifer test	Eastern Morris County	Gill and Vecchioli (1963)
0.18	"	"	"

[1]Shown on figure IV-1.

of 10^{-3} to 10^{-2} probably indicate semiconfined aquifers overlain by a leaky confining layer of lake-bottom fine sand to silt. Storativities on the order of 10^{-1} probably indicate unconfined aquifers in fluvial or deltaic sand and gravel deposits.

GROUND-WATER FLOW AND RECHARGE

Flow Paths

Ground water, under natural conditions, moves along flow paths that are determined primarily by topography and by the distribution of permeable and impermeable geologic materials. In glacial aquifers the chief controls on natural flow paths are the shape of the bedrock valley or trough containing the aquifer and the distribution and continuity of permeable and impermeable glacial sediment. Accurate determination of flow paths requires water-level readings from numerous monitoring wells completed at various levels in the aquifer system and measured regularly for months or years. Although such intensive data collection has only recently begun in a few glacial aquifers in New Jersey (Hill and others, 1992; Hoffman and Quinlan, 1994; Schaefer and others, 1993; Nicholson and others, 1996), available data allow some preliminary conclusions.

Unconfined aquifers generally have water tables that conform to modern valley topography and stream drainage. Flow paths generally follow the same direction as surface streamflow, and have a slight component away from the valley walls and toward the main stream, except where artificially deflected toward large-capacity wells or where streams are losing water to the aquifer.

Where not affected by pumping wells, flow paths in confined aquifers tend to follow the axis of the buried bedrock valley or trough containing the aquifer. In several places in New Jersey these buried valleys cross modern drainage divides, and water in the confined aquifer flows in a different, and in some places, opposite, direction from that followed by surface water or water in the unconfined aquifer. Examples include: (1) northward ground-water flow along the axis of buried valleys trending north from Budd Lake and Green Pond, both of which have surface drainage to the south; (2) eastward ground-water flow in the confined Troy Hills valley-fill aquifer (fig. IV-1) through the buried preglacial Rockaway valley, whereas the modern Rockaway River and water in the unconfined aquifer flow north (Schaefer and others, 1993); (3) probable northward ground-water flow in the confined zone of the Lamington valley-fill aquifer (fig. IV-1), along the axis of a buried valley, although modern surface drainage is to the south (this northward flow is now disrupted by pumping [Nicholson and others, 1996]); and (4) possible eastward ground-water flow in the buried preglacial Rockaway valley at Dover, where the modern Rockaway was shifted southward by deposition of the terminal moraine (Dysart and Rheaume, 1999).

In places where large closed basins have been excavated in bedrock by glacial erosion, for example, in the central Passaic River basin and Hackensack valley, water in the confined valley-fill aquifer cannot drain to lower elevations through permeable sediment in a preglacial bedrock valley. Although detailed data on the natural piezometric surface are not available owing to the disruption of natural conditions by long-term pumping in some areas, and by the absence of water-level data in others, wells completed in the confined aquifers in these locations may flow, or, they formerly flowed, at the surface (Hoffman and Quinlan, 1994). Such conditions indicate a strong upward flow component through the overlying confining material.

Recharge

The glacial aquifers of New Jersey are recharged by: (1) direct infiltration of rain and snow on outcrops of the water-producing beds, (2) direct infiltration of water from lakes, streams, and wetlands into outcrops of the water-producing beds, particularly where induced by heavy pumpage of the aquifer, and (3) flow of ground water into the aquifers from adjoining bedrock and glacial sediment.

The sources and flow paths of recharge for a given aquifer are difficult to document with certainty. Many observation wells would be needed to establish both horizontal and vertical ground-water flow paths. Aquifer tests would be required to determine the water-transmitting properties of the materials and continuity of aquifers. Stream discharge would have to be measured at many locations to determine the location and quantity of seepage into or out of the stream. Furthermore, conditions change seasonally, so these data would have to be collected over a period of months or years. Such intensive, long-term, aquifer-wide data collection has only recently begun for a few glacial aquifers in New Jersey (Hill and others, 1992; Schaefer and others, 1993; Hoffman and Quinlan, 1994; Nicholson and others, 1996). Preliminary results of these studies, and of several computer models of glacial aquifers (Meisler, 1976; Hutchinson, 1981; Hill, 1985; Sirois, 1986; Hill and others, 1992; Nicholson and others, 1996) permit some tentative conclusions.

Direct precipitation is probably a significant source of recharge for aquifers with large outcrop areas of permeable sediment. Examples of such large outcrop areas include broad glaciofluvial plains--for example, the Plainfield outwash plain east of Somerville (fig. II-3)--and broad deltas in large glacial lake basins. An example of the latter is the large delta in glacial Lake Passaic at and southeast of Morristown (fig. II-3). Hoffman and Stone (1991) consider this delta to be an important recharge area for confined aquifers to the north and east. Other examples include the Bridgeville, Germany Flats, Lamington, Ramsey, Franklin Lakes, and northernmost part of the Hackensack, valley-fill aquifers (fig. IV-1), all of which have large outcrop areas of permeable sediment.

Infiltration of surface water from streams, lakes, and wetlands is a primary source of recharge for aquifers with relatively small outcrop areas in narrow valleys with large streams, and for aquifers in contact with large lakes. Examples include the Columbia, Upper Delaware, Flatbrook, Musconetcong, Budd Lake, Netcong, Berkshire Valley, Rockaway, Ramapo, Wanaque, and Greenwood Lake valley-fill aquifers (fig. IV-1). Field studies and computer models in the Ramapo, Rockaway, and Lamington valleys (Vecchioli and Miller, 1973; Hill, 1985; Sirois, 1986; Hill and others, 1992; Schaefer and others, 1993; Nicholson and others, 1996; Dysart and Rheaume, 1999) indicate that seepage from some reaches of the trunk stream in the valley, and also from tributary streams draining the surrounding uplands, is a significant source of recharge. Seepage may be especially important where the streams cross outcrops of permeable sediment or flow near pumping centers.

Few data are available to document flow of water to glacial aquifers from adjoining bedrock and glacial sediment. Aquifer tests (Vecchioli and Miller, 1973; Canace and others, 1983) indicate that semi-confined aquifers are readily recharged through overlying leaky confining beds of fine-sand-to-silt lake-bottom sediment. Recharge through less-permeable clay and silt confining beds is minimal (Hill and others, 1992).

Aquifer tests also indicate good hydraulic connection between permeable rocks such as carbonate rock, fractured sandstone, and shale; and permeable glacial sediment (Herpers and Barksdale, 1951; Hutchinson, 1981). Recharge through bedrock may be significant where the glacial aquifers are bordered by upland areas of outcropping permeable rock, especially where the water-producing beds are confined by thick, continuous, fine-grained lake-bottom sediment. Examples include the Hackensack, Overpeck, Lower Passaic, and Kenilworth valley-fill aquifers (fig. IV-1), which are bordered by uplands formed on permeable sandstone and shale of the Passaic Formation, and are locally confined by thick lake-bottom silt and clay. Similarly, parts of the Pequest, Germany Flats, Vernon Valley, Wallkill, and Lafayette valley-fill aquifers (fig.IV-1) are bordered, at least on one side, by low uplands on permeable carbonate rock and are also locally confined by lake-bottom sediment. In these locations, infiltration of precipitation and of water from lakes, streams, and wetlands into the bedrock on the uplands may be a significant source of recharge for the confined glacial aquifers.

CONCLUSIONS

The principal glacial aquifers in New Jersey are stratified valley-fill deposits greater than 100 feet thick. They are chiefly of late Wisconsinan age but are of Illinoian age in a few places.

Several types of valley-fill aquifer are distinguished based on the environment of deposition. Glacial lake basins that did not fill completely with sediment may contain a basal confined aquifer of sand and gravel deposited in lacustrine fans, overlain by silt and clay lake-bottom sediment. Sand and gravel in deltas and outcropping lacustrine fans may form unconfined aquifers. Glacial lake basins that were completely filled with sediment include these deposits plus an upper sand and gravel deposited by streams flowing on the former lake bottom. This upper sand and gravel may form an unconfined aquifer. Valleys that were not occupied by glacial lakes contain sand and gravel deposited by glacial streams. Because these valleys do not have extensive deposits of silt and clay, ground water is under unconfined conditions. Finally, several valleys along the terminal moraine of the late Wisconsinan glacier contain a more complex fill that includes lacustrine sediment and till of Illinoian age overlain by similar sediment of late Wisconsinan age. In these valleys there may be several water-bearing beds and several confining beds.

Published yields of 317 large-capacity wells tapping glacial deposits range from 6 to 2200 gallons per minute (gpm). Specific capacities of these wells range from 0.19 to 500 gpm per foot of drawdown. Hydraulic conductivities of aquifers, determined from 47 tests, range from 0.0014 to 0.026 foot per second (ft/s). Hydraulic conductivities of confining beds are 4 to 8 orders of magnitude smaller. Aquifer storativity values range from 10^{-4} to 10^{-1}.

Identification of flow paths and recharge areas equires detailed study on a valley-by-valley basis. Existing data suggest that ground-water flow in confined aquifers is controlled, in part, by the topography of the buried bedrock valley or trough containing the aquifer. Flow in unconfined aquifers generally is in the direction of modern stream drainage. Recharge may occur through both outcropping parts of water-bearing beds and through bedrock and glacial deposits adjoining the aquifers.

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CHAPTER V

Water Table Gradients in Unconsolidated Deposits at Small Contaminated Sites

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ABSTRACT

Water table gradients at most contaminated sites regulated by the New Jersey Department of Environmental Protection, Bureau of Underground Storage Tanks are determined from a minimum of three wells screened in the same water bearing zone. At most sites, the water table gradient and the inferred directions of ground water flow and contaminant transport are in the same direction as topographic slope. At a substantial minority of sites, however, gradients are upslope, across slope, unreasonably steep, radially inward, or in some other way appear to be anomalous. The anomalies can commonly be explained as the result of geologic conditions or human activities. This presentation offers case studies of "normal" gradients (at a reasonable steepness in the direction of topographic slope) and water table or contaminant concentration gradients which are not in the direction of topographic slope because of downward components, horizontal components of percolation above the water table, perching, mounding beneath permeable fill, and drawdown by offsite pumping.

INTRODUCTION

Where contamination at the water table is a concern, New Jersey Department of Environmental Protection regulations require determination of ground water gradient as part of an initial investigation. The direction and steepness of the gradient is determined by measurement of water levels in at least three monitoring wells. The wellheads are surveyed vertically to within 0.01 foot, and ground water elevations are determined from depth to water. The gradient and presumed direction of ground water flow and contaminant transport are then inferred from the gradient. At most sites, the water table gradient is in the direction of the topographic slope at a few percent or less (figure 1). At some sites, apparent gradients in directions other than the topographic slope are the result of errors in measuring water levels, errors in surveying of well elevations, or wells being screened in more than one water-bearing zone. At other sites gradients in directions other than the topographic slope have been confirmed as accurately reflecting water table conditions by confirmatory surveying, persistence through several monitoring rounds, consistency with contaminant concentration gradients, installation of additional monitoring wells, or other lines of evidence.

This presentation offers field examples of "normal" flow (ground water and contaminant transport in the same direction as the topographic gradient) and of apparently anomalous flow, not as would be predicted from topography. The examples are from New Jersey Department of Environmental Protection, Bureau of Underground Storage Tanks cases. Seemingly anomalous flow is explained as the result of heterogeneous subsurface conditions, vertical components of ground water flow, or human intervention.

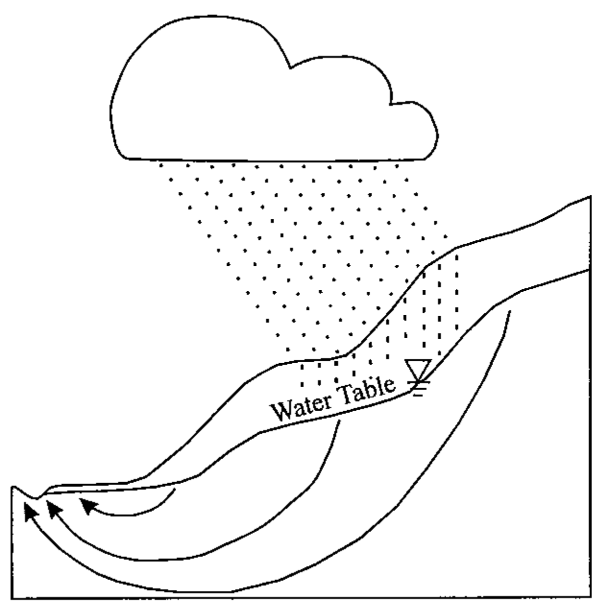


Figure 1. Ground water originating from precipitation, percolating vertically to the water table, then migrating in a downslope direction to a discharge area.

Example 1. Downslope Water Table Gradient and Contaminant Transport

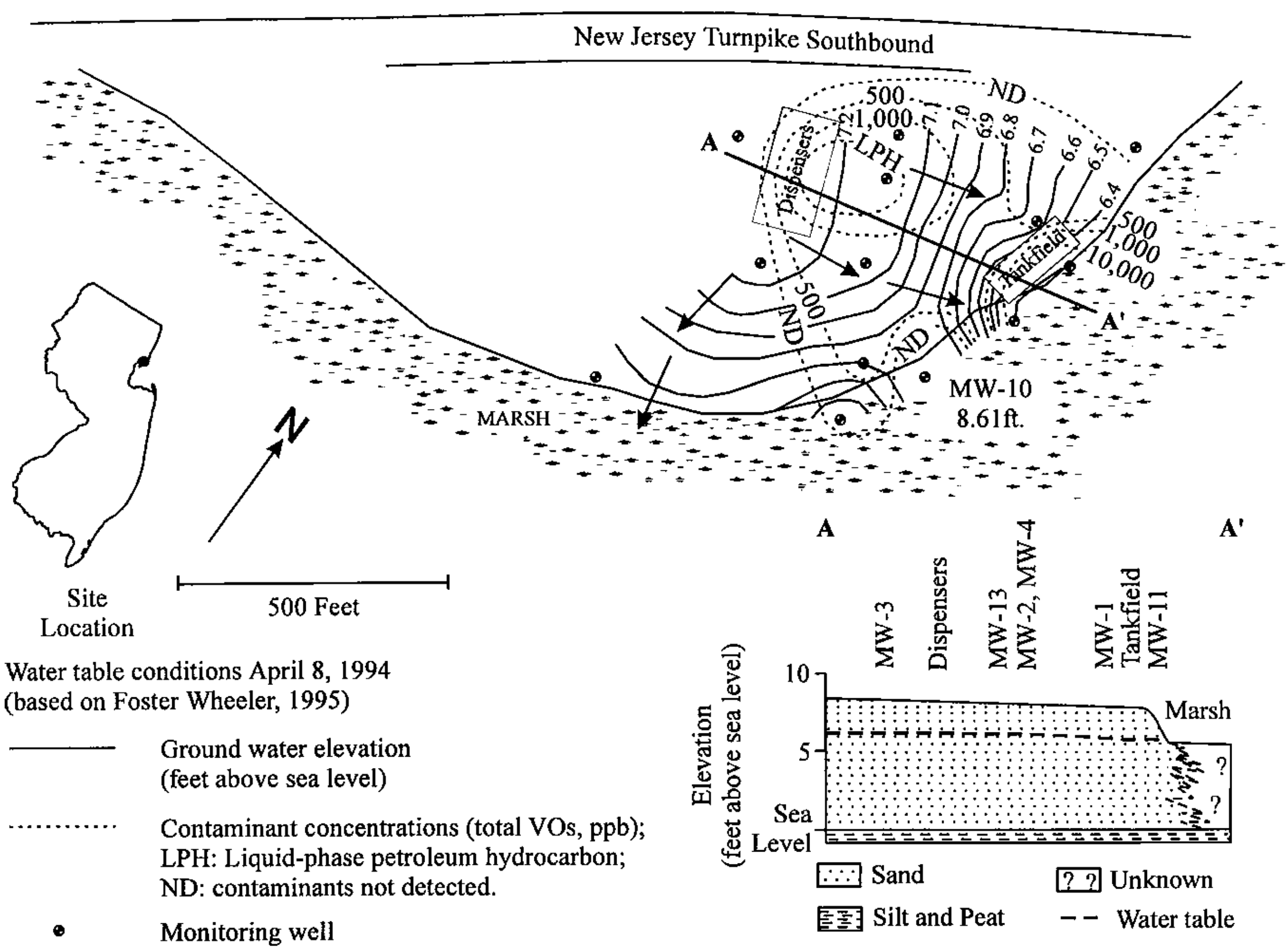
This case involves an area of sandy fill built outward from the New Jersey Turnpike for a rest area. The fill is built into marsh to an elevation of about eight feet (about three feet feet above the level of the marsh). Ground water is under water table conditions and is confined below by silt. There is no evidence from well logs or excavation descriptions of enough heterogeneity to cause significant perching. There is no reason to predict vertical flow upward or downward through the underlying silt. The rest area was closed in 1978, and all gasoline handling equipment was removed. It still remains contaminated by gasoline related compounds.

As would be predicted from topography, gradient is radially outward. As would be expected from the permeable nature of the fill, the gradient is gentle and the water table does not rise much above the marsh. The gradient is almost horizontal towards the center of the site and increases to about 0.002 feet per foot towards the discharge area at the marsh. The water table at the center of the site is about 0.8 feet above the marsh.

Based on concentrations, contamination appears to be moving downgradient from two source areas, the former dispensers and the former tankfield. Around the dispensers, wells immediately upgradient and sidegradient to the edge of the marsh are not contaminated. The nearest downgradient well, 150 feet from the dispensers, was contaminated with liquid phase hydrocarbons as of April 8, 1994. Wells to 250 feet downgradient exceeded 500 parts per billion (ppb) total volatile organic compounds (VOs).

Movement of contamination from the tankfield is not as well documented because wells have not been installed beyond the edge of the marsh, but contamination appears to be moving downgradient from the tankfield as well as from the dispensers. The one immediately upgradient well shows only a few hundred ppb VOs, probably from the dispensers. The two downgradient wells show over 10,000 ppb VOs.

Ground water elevation and contaminant concentration at well MW-10, at the edge of the marsh, is anomalous. Even though MW-10 is adjacent to the presumed discharge area, it has the highest water level of any well at the site, about 1.5 feet above the water level in the wells toward the center of the site. Even though the well is surrounded on three side by contaminated wells, VOs were not detected. A preliminary interpretation of this anomaly is that water levels are kept high and contaminants are kept low by a discharge of clean water, possibly from a storm drain outlet or leaking water pipe. This is similar to the ground water mounding effects described in example 8.



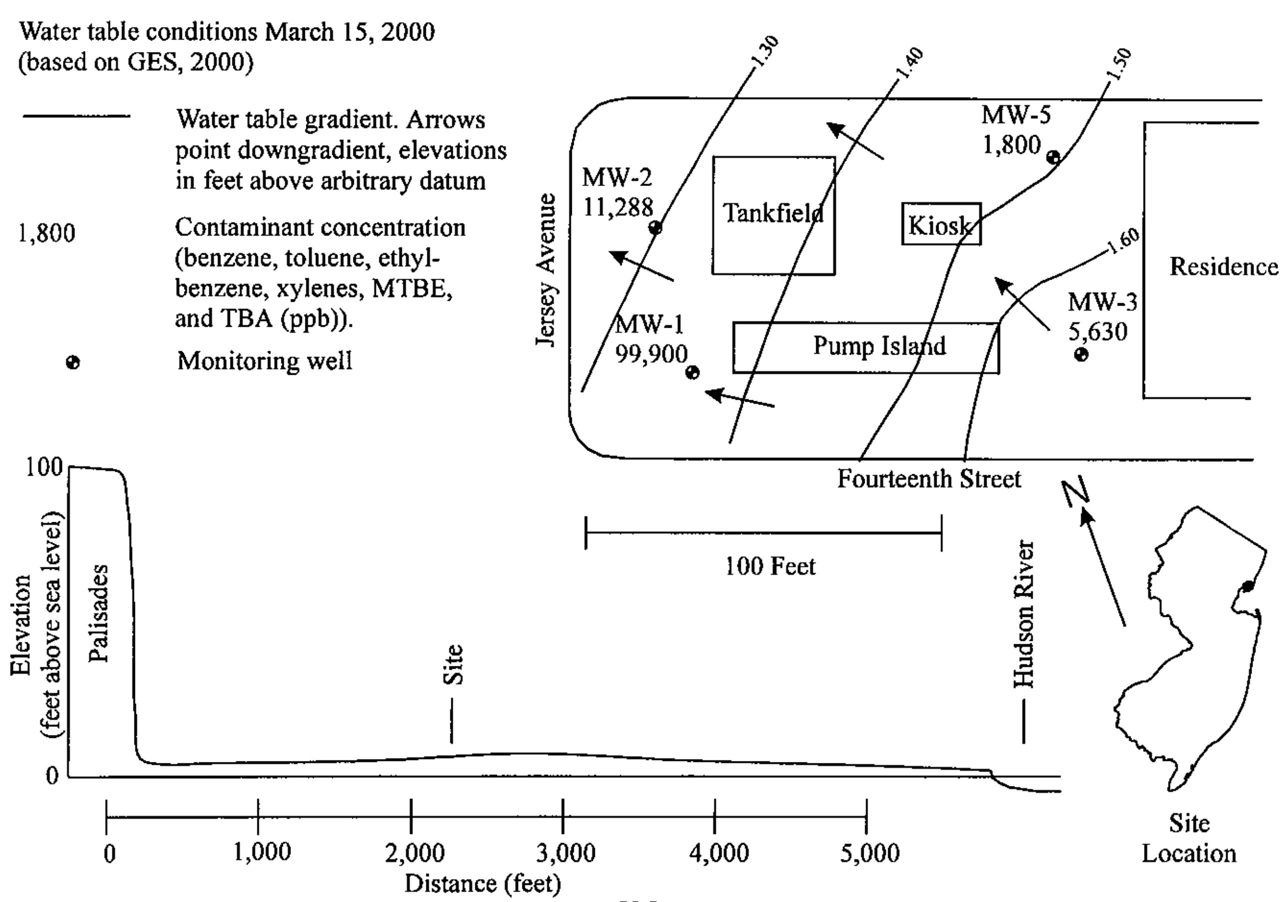
Water table conditions April 8, 1994 (based on Foster Wheeler, 1995)

Example 2. Ground Water Flow Following the Most Local Topographic Gradient Discernible from 1:24,000-scale Topographic Maps

At most sites in New Jersey, the water table is between 5 and 20 feet below ground surface. When ground water is this close to the surface, gradient usually parallels the most local topographic gradient discernible on 1:24,000 scale U.S.G.S. topographic maps. Gradients are towards small swales and dips, even when the overall topography and the nearest surface water shown on topopgraphic maps are in a different direction. This site, in Jersey City along a connecting road between the Holland Tunnel and the New Jersey Turnpike, illustrates such a situation. The site is in filled land at an elevation between 10 and 20 feet above sea level. The Palisades rise to an elevation of 100 feet about 2,000 feet to the west of the site. The Hudson River is at sea level 3,500 feet to the east. The immediate area is almost level, but the highway configuration between the Holland Tunnel and the Turnpike on-ramp creates an impression that the topographic gradient is to the east, towards the Hudson River. Identification of the direction of contaminant transport is important at this location because of high concentrations of volatile organic compounds in MW-1 and in soil and the consequent need to evaluate vapor hazards in the basement of an adjacent residence. U.S.G.S. topographic maps contradict the surface impression that there is an eastward slope. There is instead a very gentle westward slope towards railroad yards at the base of the Palisades. No surface drainage along this low area is evident from topographic maps or from surface inspection.

Ground water gradients are persistently westward, away from the residence (towards the Palisades and away from the Hudson). Although contamination is not well delineated, concentrations of volatile organic compounds are consistent with the water table gradient in being higher downgradient from likely sources of contamination at the pump island and tankfield. Air monitoring in the basement of the residence has shown no readings above background.

The pattern of shallow ground water flow towards a minor topographic feature which might not be perceived as controlling subsurface conditions is common and appears to reflect a more intimate relationship between topography and ground water gradient than is apparent from surface water drainage. As in many cases in densely developed urban areas, however, there are alternative explanations for the gradient at this site. The water table is only 6 to 8 feet below ground surface. Flow could, for example, be towards deeply buried utility lines, tunnels, or drainage sumps in nearby buildings.



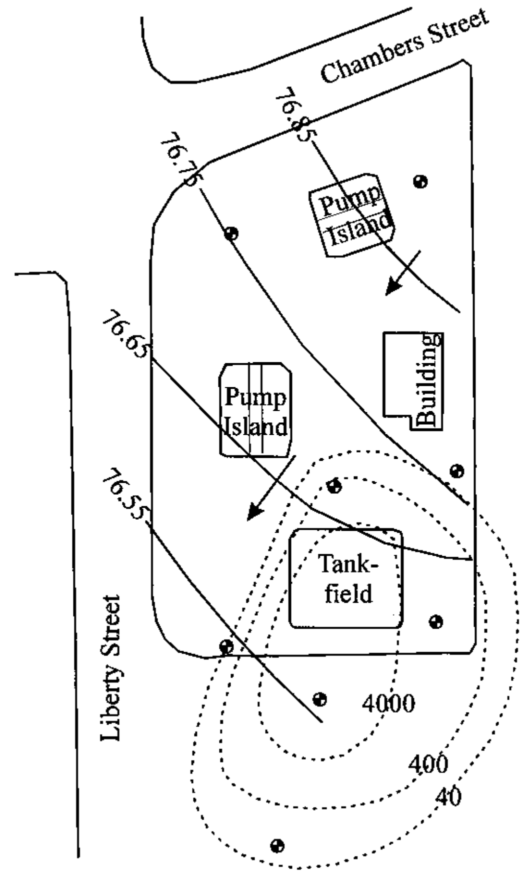
Example 3. Ground Water Gradient Towards a Regional Base Level

Where there is a substantial thickness of permeable sediment above the ground water base level, the water table in New Jersey may be 40 feet or more below ground surface, well below small-scale topographic irregularities. As would be predicted, ground water this deep does not respond to small topographic irregularities as it does at sites with shallower ground water. Instead it follows a more regional gradient through the permeable sediment towards the nearest surface water body.

At the Hamilton Township, Mercer County gas station illustrated here, well logs show sand and gravel of the Pensauken Formation and possibly the underlying Raritan-Magothy aquifer system to the 40 foot depth of drilling. Ground water is at 25 feet. The topography is essentially flat for 4,200 feet across a sand and gravel surface to a 50 foot escarpment overlooking a tidal section of the Delaware River. The gradient from wells at the site is 0.003 feet/foot towards the river. Projected southwestward, this would be about 12 feet above river level at the Delaware. With the expected steepening of the gradient towards the discharge area, the water table would reasonably intersect land surface at the edge of wetlands along the flood plain.

Some degree of confidence that the gradient reflects ground water flow is gained from the reasonable steepness of the gradient, persistence of the gradient since drilling of the monitoring wells in 1992, plume geometry indicating contaminant transport in the same direction as the gradient, and a similar water depth, gradient, and direction of contaminant transport at a second site 100 feet to the southeast.

As in most of the examples, the interpretation appears reasonable but is not proven and should not be generalized beyond the site. Water levels between this site and the river might, for example, be affected by perching or pumping, and nowhere near the level predicted from the projected water table.

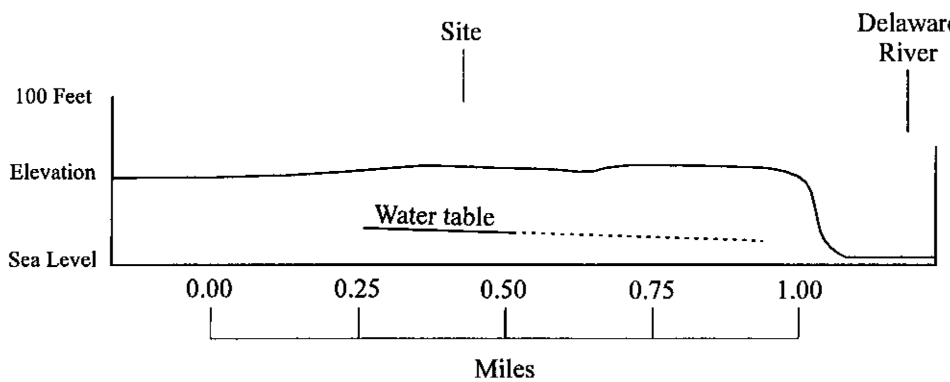


Water table conditions, September 24, 1996 (based on Hess, 1997)

Water table gradient - arrows point downgradient, elevations in feet above an arbitrary datum

Contaminant plume from tankfield (total xylene, ppb)

Monitoring well



Site Location

Example 4. Ground Water in Upper and Lower Zones Flowing towards Different Base Levels

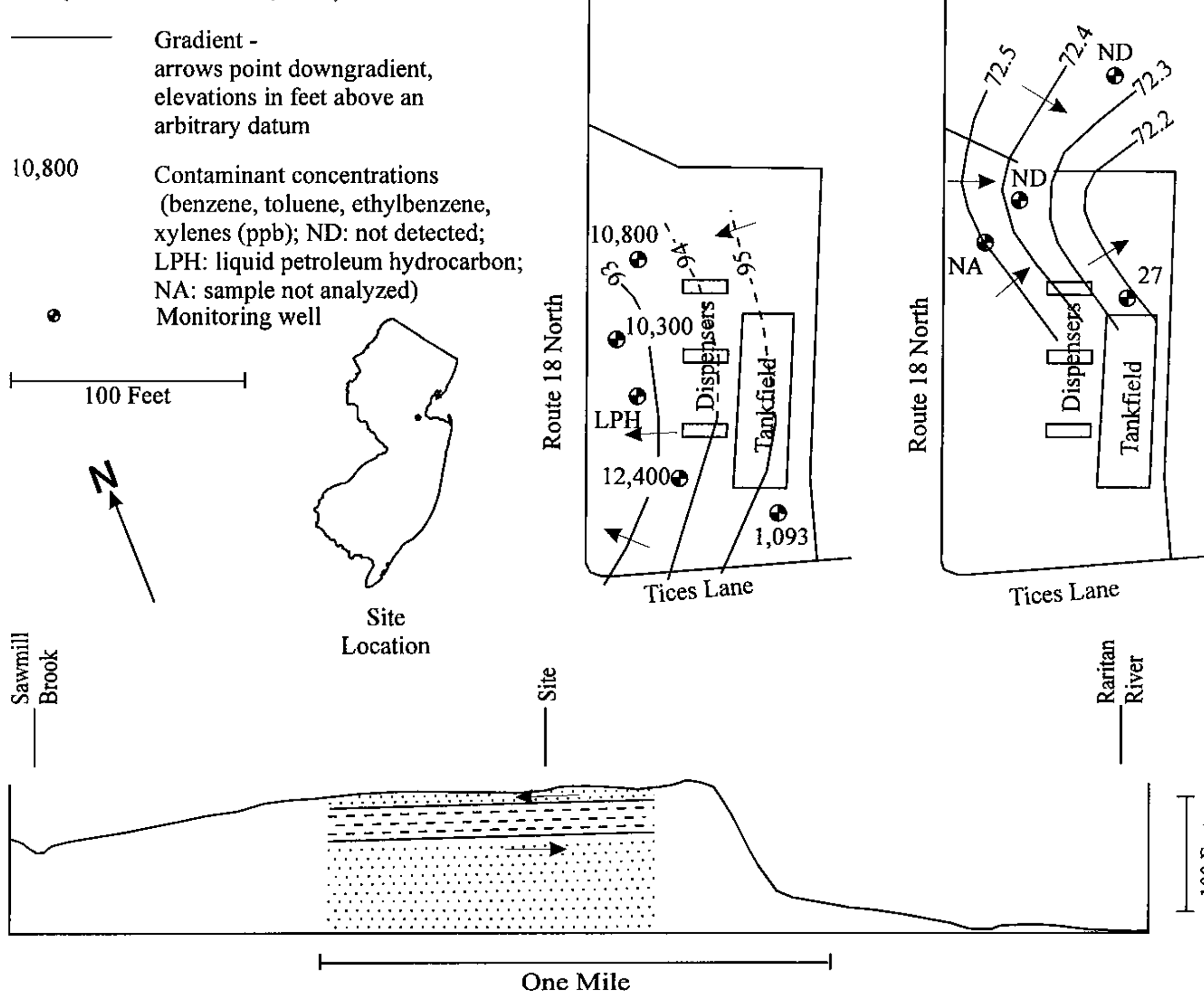
Heterogeneous sediments can exhibit complex ground water flow, but, as in the previous examples, topography may ultimately control gradients. This site in East Brunswick is near the divide between the Raritan River and Sawmill Brook. Well logs show 5 to 20 feet of sand overlying a clay layer 5 to 25 feet thick. The clay is in turn underlain by sand to an undetermined depth. The site is alongside a former clay pit, and the irregular clay thickness may be the result of clay mining followed by grading to the highway level.

Water table gradient in the upper sand is 0.04 feet/foot to the west, roughly the same direction as the topographic slope towards Sawmill Brook. Gradient in the lower sand is 0.003 feet/foot to the east. While this is opposite the topographic slope in the immediate area, it is towards the Raritan River floodplain along a flow path which is shorter and significantly steeper than towards Sawmill Brook and does not include a confining clay (see profile at bottom of page). The observed flow against the local topographic gradient at this site thus might be expected.

In the upper sand, as shown below, areas downgradient from the tankfield and dispensers were severely contaminated as of February 1992. Subsequent drilling has shown that contamination is less severe upgradient and sidegradient. Contamination throughout the site has decreased, in part as the result of active remediation.

In the lower sand, even though separate-phase hydrocarbon was present at the site for several years, ground water has remained essentially uncontaminated. Containment of contamination in the water table aquifer is common, but not universal in sites affected by gasoline-related compounds.

Ground water conditions February 7, 1992 (based on Land Tech, 1992)



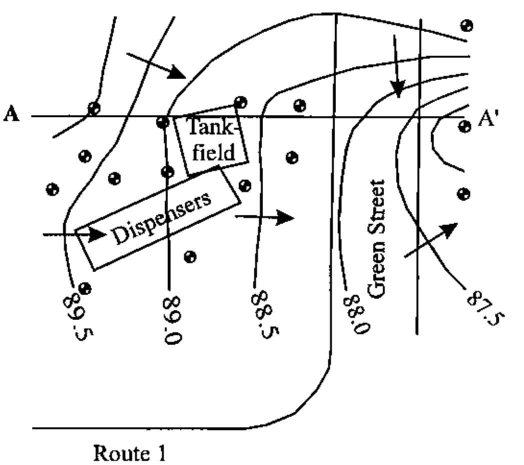
Example 5. Closed Contours Resulting from a Downward Component of Gradient

Closed contours resembling a cone of depression are common at underground storage tank sites, particularly in bedrock. This situation, in Iselin, is in glaciated coastal plain sediments. Well logs show silty sand up to 25 feet thick overlying coarser sand (cross section at lower right). The silty sand is thinnest in wells drilled near the tankfield, where it is about 12 feet thick. Depth to water is usually 10 to 12 feet. Much of the time, gradient contours are northeastward towards Woodbridge Creek. At other times the contours are inward towards areas with the least thickness of silty sand. Vapor and contaminated water have been recovered at this site, but not yet at the times shown here. The inward gradient is thus not a cone of depression from pumping. Without pumping, inward gradients are difficult to explain unless ground water flow has a downward component. Without pumping or downward flow, inward flow would eventually either fill the depression or cause ground water to rise to the land surface.

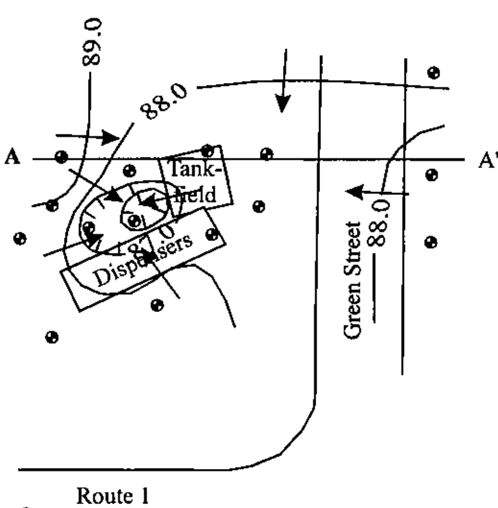
In a preliminary interpretation, gradient is taken to have a downward component at all times. At times of abundant precipitation, recharge is greater than vertical flow, and the excess recharge flows horizontally, establishing a northeastward gradient. When precipitation and recharge decrease, vertical flow becomes greater than recharge and is most rapid where the silty sediments are thinnest, only a foot or two thick below the water table, and slower where silty sediments below the water table are thicker, up to 10 feet below the water table. The relatively rapid downward flow causes a water level drop in the tankfield area, creating the inward gradient and closed contours.

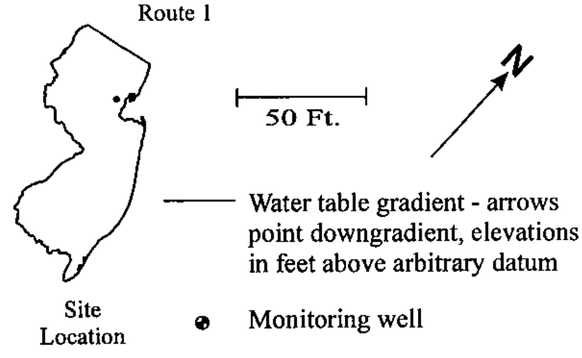
In this interpretation, the water levels are taken to be primarily the result of head in the silty sand. The wells are screened in part in the underlying sand, however, and nothing is known about the relative water levels and permeabilities of the silty sediment and underlying cleaner sand. If the water levels reflect head in the sand rather than in the silt, the closed contours would be difficult to explain as the result of the thin silt near the tankfield. Interpretation at many sites must be made under similar conditions of uncertainty, and intuitive reasoning commonly provides the best interpretation possible.

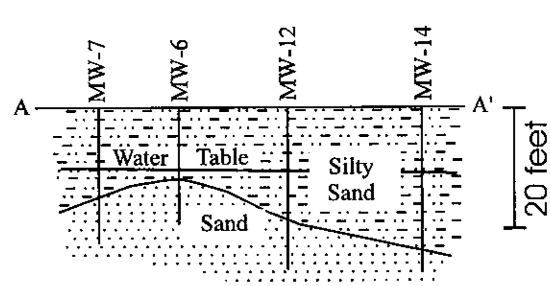
Northeastward gradient March 1, 1994 (based on IT, 1994)



Inward gradient, December 1, 1995 (based on IT, 1995)







Example 6. Apparent Steep Contours Resulting from Perching

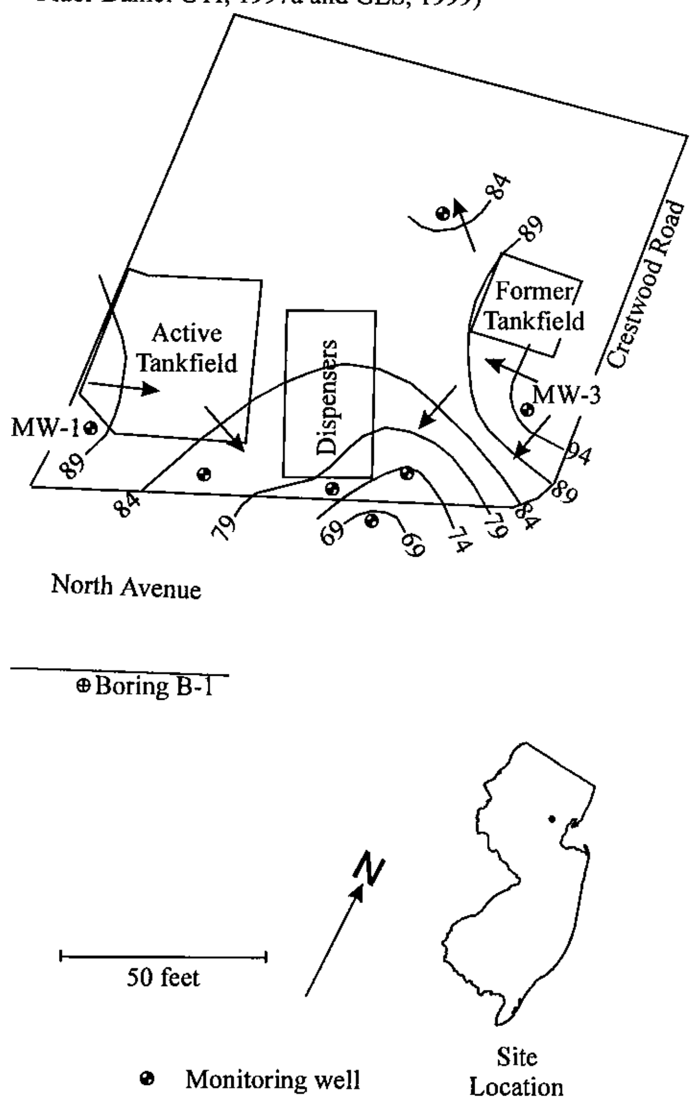
At some sites local near-surface perching creates the appearance of steep vertical gradients. This site is on till near the crest of the terminal moraine in a commercial area of Scotch Plains. Water levels appear to show gradients at about 30 percent to the south. While the gradients have been consistent since the installation of monitoring wells in 1990, a 30 percent gradient is not reasonable. A 25-foot drop in ground water elevation across an 85 foot distance does not call to mind a normally sloping ground water gradient under New Jersey conditions. At this gradient, ground water would be at sea level within 700 feet of the site.

The most reasonable interpretation for these water levels is local perching at depths of 10 feet or less. Till in this area is sandy and appears to be permeable. The terminal moraine is elevated above areas to the north and south, and depth to ground water in the permeable till is deeper than usual in New Jersey. Boring B-1, for example, (across North Avenue from the site) was 60 feet deep and was dry. The log shows medium to coarse clayey sand and gravel overlying weathered shale. The shale is at 57 feet.

While the lithologic logs of the monitoring wells do not confirm silty perching layers, the behavior of monitoring well MW-3 offers speculative insight into the nature of the near-surface conditions. This well is 20 feet deep and is screened from 3 to 20 feet. Depth to water in about half of 23 sampling rounds was between 1 and 5 feet. Depth to water was between 16 and 19 feet on three rounds, and the well was dry on 10 sampling rounds. All of the sampling rounds in which depth to water was between 1 and 5 feet were between January and April, months generally associated in New Jersey with high recharge and low evapotranspiration. All of the sampling rounds in which depth to water was 16 feet or greater were between June and December, months in which evapotranspiration is high or ground water is recovering from lowering through the summer months. Based on this, it appears reasonable that perching occurs at MW-3 at a depth of 5 feet or greater through periods of high recharge and low evapotranspiration. The perching layer is not substantial enough, however, to maintain ground water through the summer months, however, and the well goes dry.

A similar situation can be inferred from water levels in MW-1, near the active tankfield. All 23 monitoring rounds except 3 show ground water elevations higher than 10 feet below ground surface. The three rounds showing depth to water greater than 10 feet are between September and November, following seasons of high evapotranspiration and before water levels recover to their seasonal maximum towards the end of winter or early spring. While the perching at MW-1 appears to be more effective, ground water behavior at both MW-1 and MW-3 is consistent with the seasonal perching of ground water at depths of roughly 10 feet. Perching may occur naturally above a fine-grained layer within the till. Alternatively, as the wells affected by apparent perching are in the near vicinity of the active and abandoned tankfields, perching may occur at the base of permeable fill within tankfield excavations.

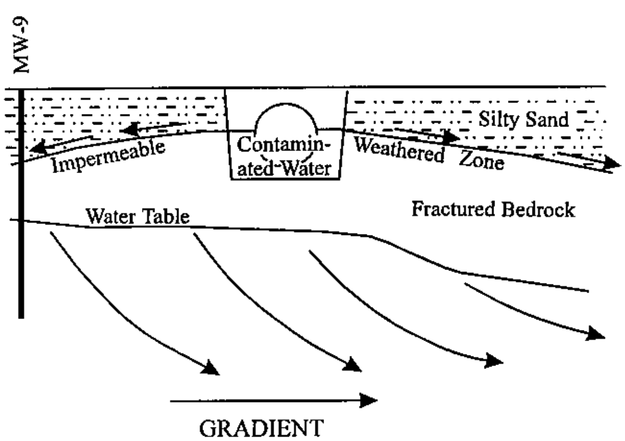
Apparent water table gradient, March 1, 1996 (based on Fluor Daniel GTI, 1997a and GES, 1999)



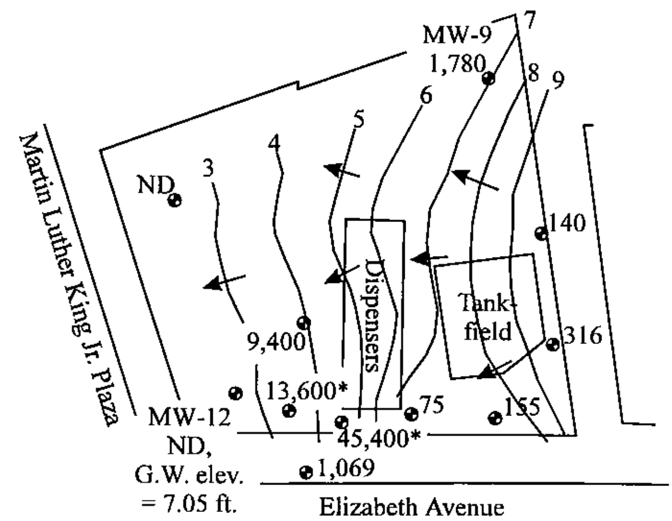
Example 7. Apparent Sidegradient Contaminant Transport Resulting from a Horizontal Component of Percolation above the Water Table

Percolation to the water table is usually assumed to be vertical, but may have a horizontal component if, for example, sloping fractures, sandy layers, or relatively impermeable zones are present. This site in Elizabeth is underlain by shale of the Passaic Formation. As in many sites in the Newark Basin, a few feet of relatively permeable sediment (in this case silty sand, probably loess or glacial lake sediment) overlies shale weathered to a relatively impermeable clayey consistency. The weathered zone is a few feet thick above fractured, more permeable bedrock. While the weathered shale is less permeable than the overlying sand or underlying fractured bedrock, it does not support a persistent perched water table. The water table is in bedrock several feet below the weathered zone.

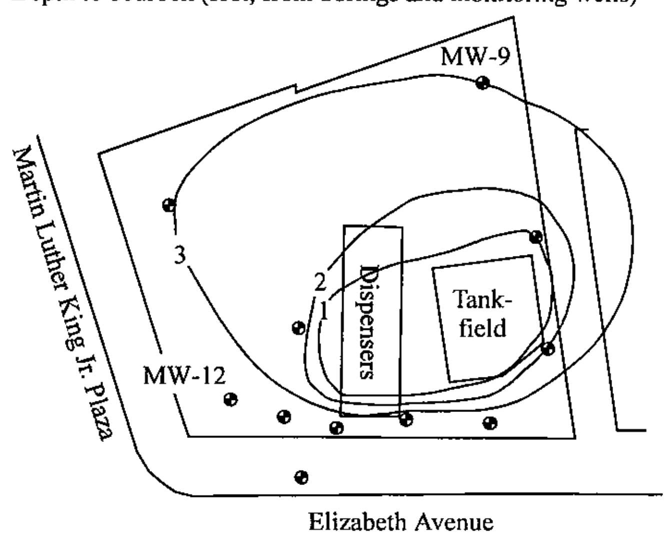
A westward water table gradient is consistent with topography in being downslope towards a swale leading to the Elizabeth River. Downgradient contaminant transport is generally confirmed by higher concentrations in wells downgradient to the west of the tankfield and dispensers. Contaminated well MW-9, however, is roughly 60 feet sidegradient from any obvious source of gasoline-related compounds. Contamination at this well is best explained by a horizontal component of transport at the top the weathered zone. Wells and soil borings show that the tankfield is dug into shale at the top of a bedrock knoll which has been graded over. While the weathered zone does not create a persistent perched water table, contaminated water in the tankfield is interpreted here as rising above the weathered bedrock at times of high recharge, and percolating along the top of the weathered zone towards MW-9 (diagram below).

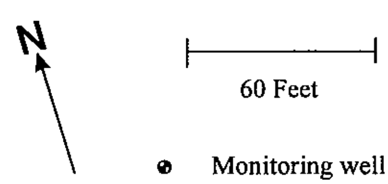


Water table elevations, June 7, 1993 (feet above arbitrary datum)
Contaminant concentrations, June 28, 1993 (ppb, total of benzene, toluene, ethylbenzene and xylenes; *well not sampled June 28, 1993; value is for closest sampling date)



Depth to bedrock (feet, from borings and monitoring wells)





Based on Fugro McClelland, 1994

Site Location

Examples 1 through 7 illustrate natural conditions. Human intervention in the water cycle also creates gradients which would not be predicted from topography. Examples 8 through 10 show what may be the three most common ways human activities influence gradients and contaminant distributions at underground storage tank sites in New Jersey: mounding beneath permeable fill, flow along trench fill, and water table lowering by pumping.

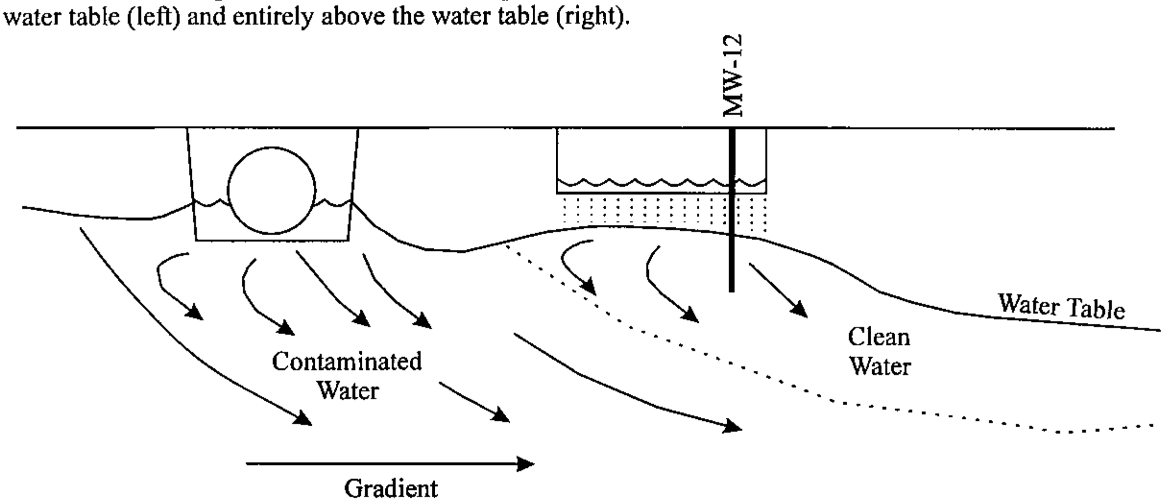
Example 8. Ground Water Mounding

Backfill at underground storage tanks is commonly more permeable than the native soil. Pea gravel, for example, is a common backfill. Water collects within the fill and infiltrates constantly rather than only during and after precipitation or snowmelt. Increased total recharge beneath the permeable material results in ground water mounding. Mounding is common at underground storage tank sites whether the permeable fill intersects the water table or is entirely above it (diagram below).

Increased recharge below excavations can lead to higher or lower contaminant concentrations as well as higher water levels. Infiltration through a gravel-filled excavation may feed large volumes of either contaminated or clean water into the ground. The same Elizabeth gas station used in example 7 to illustrate flow at the top of a weathered zone offers an example of "clean" water infiltrating permeable fill and creating a mound of uncontaminated water within a contaminant plume. Monitoring well MW-12 is downgradient from the gasoline dispensers and is close downgradient from three severely contaminated wells. Even though the well is in a downgradient location, the water level in the MW-12 is as much as four feet above the levels in the nearby wells. Water samples consistently test "clean" for gasoline-related volatile organic compounds. Well records provide a likely explanation for this "clean" well so close downgradient from contaminated wells. Logs of most wells at the site show a few feet of silty sand overlying bedrock. MW-12 shows clayey silt with medium to fine gravel to 8 feet. Water saturated timbers were noted from four to six feet. The well is interpreted as drilled through fill in an abandoned excavation, possible the basement of a building formerly at the site. The "clean" water is interpreted as a lens created by increased infiltration of water unrelated to the source of contamination. Based on this, water samples from MW-12 were judged to be unrepresentative of the plume and of little value in evaluating the progress of remediation. Sampling was discontinued, and the well was sealed.

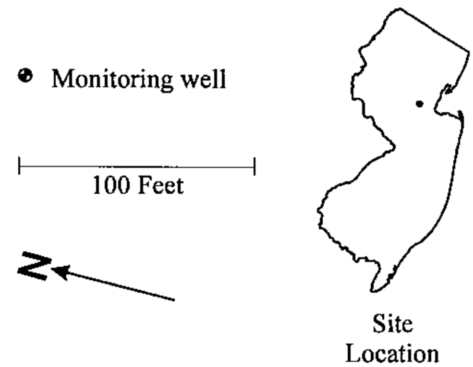
While mounding is most common beneath fill, it is the result of increased recharge and not necessarily related to fill. A similar "clean" well within a plume was previously noted at the abandoned Turnpike rest area (MW-10 in the first example). Permeable fill is unlikely as the cause of this mounding as the entire site is underlain by permeable fill. As mentioned in that example, discharge from a storm drain or leaking pipe is the more likely explanation for the mounding at the Turnpike rest area.

Ground water mounding beneath excavations completed into the water table (left) and entirely above the water table (right).

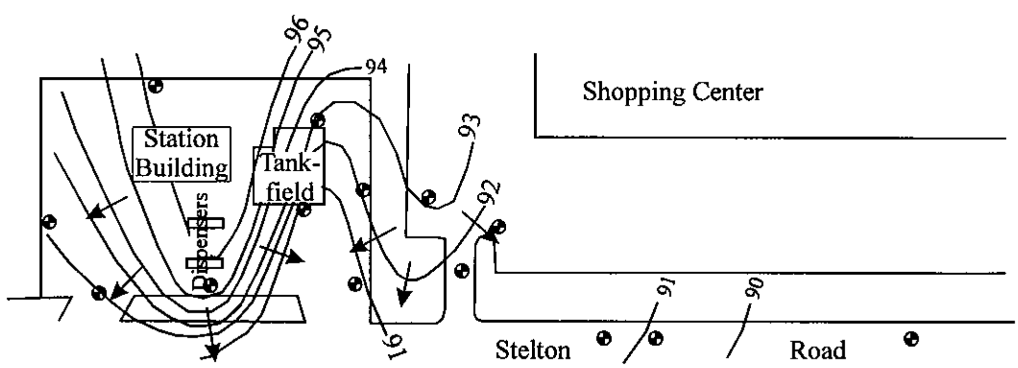


Example 9. Ground Water Movement along a Trench

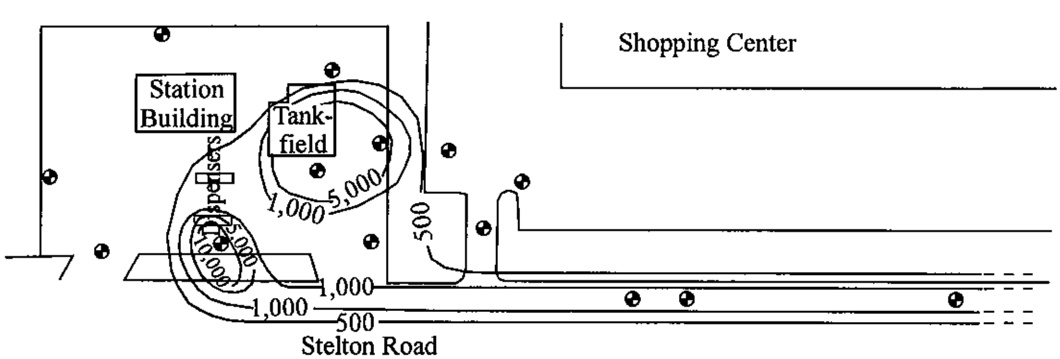
One of the most common activities to affect ground water flow at contaminated sites is trenching. Whether a trench is above or below the water table, water and contaminants can move along permeable fill in directions unrelated to surface topography. This example is from a gas station on shale of the Passaic Formation in Piscataway, Middlesex County. Topographic slope, water table gradient, and contaminant transport are westward towards Ambrose Brook. Separate contaminant plumes originate at the tankfield and the dispenser area (lower diagram). At Stelton Road, the contaminant plume takes a right angle bend and moves southward through utility trench backfill. Ground water in the trench is usually two to three feet below ground surface. Absence of further westward spread of the contamination is confirmed by "clean" monitoring wells across Stelton Road from the site. Apparently, even though the trench is dug no more than few feet into the water table, the bulk of the contamination is so shallow that the trench acts as a barrier to downgradient contamination movement.



Water table gradient, July 7, 1993, elevations in feet above an arbitrary datum, arrows point downgradient (based on Fugro McClelland, 1994)

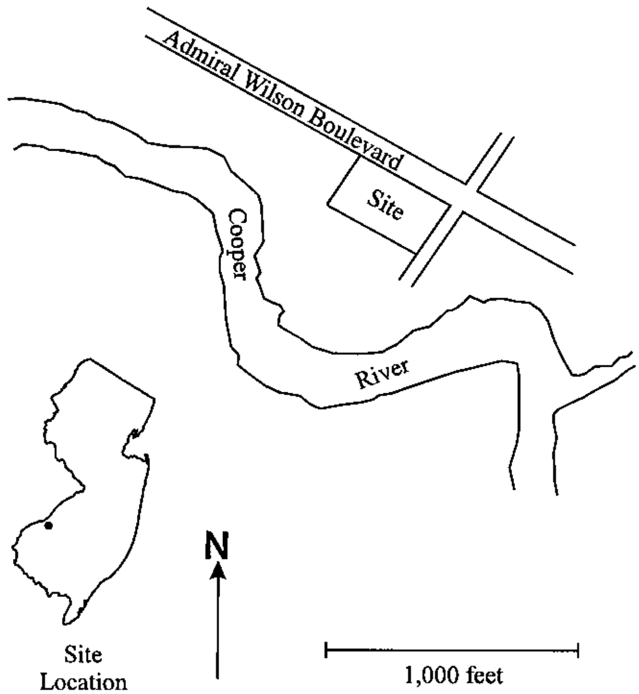


Contaminant concentrations April 3, 1991 (total of benzene, toluene, ethylbenzene xylenes (ppb); "clean" offsite wells to west across Stelton Road not shown



Example 10. Water Table Lowering by Pumping

After mounding beneath filled areas and migration along trenches, the most common human intervention affecting water tables at New Jersey underground storage tank sites may be pumping. In this example, contamination was discovered at a site along the border between Camden and Pennsauken about 750 feet northeast of the Cooper River. The site is on the flood plain about 10 feet above river level. Well logs show sand with subordinate silt and gravel to the 50 foot depth of drilling. During an initial investigation in 1989, ground water was at about 40 feet, 30 feet below river level. Subsequently, ground water levels rose to about 33 feet in February 1995, and appear to have stabilized at about 25 feet from 1997 to 1999. Gradients have remained away from the river at about 0.002 feet/foot through the period. Based on discussions with U.S. Geological Survey - Water Resources Division personnel (LaCombe, personal communication), withdrawals from Camden area well fields were cut back through this period as the result of contamination. Additional regional cutbacks were begun in 1996 as part of a water supply management effort to alleviate overuse of the Potomac-Raritan-Magothy aquifer system in the Camden area. The rising water levels are attributed to water level recovery following these cutbacks. The rising ground water is a concern as petroleum saturated soil may now be submerged 15 feet below the water table and may be more difficult to locate and remediate as a result.



Based on Fluor Daniel GTI, 1997b

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CHAPTER VI

MENTORING THE BADD WAY (WORKSHOP)

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The worlds of academia and industry have the opportunity to meld for the benefit of a student. Last school year and again this year, we were given the opportunity to act as mentors to a high school senior. The placement found us both very excited, but slightly apprehensive, due to our previous experiences in cooperative education programs. Both of us attended the same college a mere twelve years apart and had attempted to enter a similar program. One of us was sent to a regulatory agency where the interviewer was not prepared and did not listen to the student's wants or needs. The other was inappropriately placed and missed out on potential funding and a better placement when the cooperative education counselor did not respond to requests for information. One of us ended up not interning and the other was placed in a field without the necessary educational background.

However, the high school that we worked with had a terrific program in place. The mentorship program places a senior in a science or engineering firm for a semester (with the possibility of an extension). Students spend two mornings a week at their placement instead of attending community college classes as required by the school. The mentor is given latitude on the assignments he or she gives the student. The student may work on multiple or individual projects as part of a team based upon discussions between the mentor, student and the school. Some students are given actual projects as our intern was, while others are given fictional projects to work on. The research effort must involve data and information gathering necessary to conduct any scientific or engineering endeavor. Activities such as library reviews, preparing tables and graphs, and internet searches are a part of the student's experience. The mentors are informed that the student assigned to them is more than likely not experienced and will need some guidance. However, the school asserts that the student should be given responsibility and independence.

The school pairs their students with a faculty representative who acts as a liaison between the mentor and the school. The faculty representative also conducts an office visit to monitor the student's working environment. During that visit, the student, the mentor (or mentors) and the faculty representative have an informal meeting discussing the student, their work, and their progress. Additionally, the school demands punctuality from their students and that grades be maintained. If grades slip or the student is habitually late, then there is a loss of school privileges.

The students and their potential mentors meet prior to the placement to discuss potential projects and to see if the match is a good one on both sides. The mentor is asked at that time to give the student some background reading to prepare them for the placement. The student must keep a daily log, a sign in sheet and a notebook in which they collect data and information about their project. The notebook doubles as the student's resource for preparing a self evaluation paper and presentation.

The mentor should act as the student's guide as they journey throughthe work environment. By assigning work, which should stretch the abilities othe student (not beyond the students capability), then working with the student to complete their tasks (not by doing the work), benefits to both the mentor and student are acquired. These benefits are, but not limited to, the mentor getting a new outlook on the direction of his or her career and possibly new insights into his or her profession; while the student gains insights into their potential career path and invaluable experience.

When the three parties involved with the mentorship cooperate and work closely together the student gains an appreciation of the adult working world as it truly is. When they do not, the student becomes frustrated to the point of leaving the profession of their choice. Worse, they may be taught a misleading viewpoint of the professional environment which will lead to disillusionment when they enter the work force. When everything comes together, as it did with our intern, the student gains experience in the profession, the mentor gains access to a pool of experienced employees, and the school will gain a reputation of helping students not only with their education but with their careers.

CHAPTER VII

ROAD LOG AND STOP DESCRIPTIONS

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The road log starts from westbound Interstate Route 80, exit 25. Figure VII-1 shows the trip route, stops, and glacial deposits and glacial limits of the field trip area. The road log to mile 13.7 is by Scott Stanford; after that it is by Ron Witte.

Miles

0.0 Exit onto US 206 north from I-80 at interchange 25.

0.3 Exit from US 206 to International Trade Center/Waterloo Village. From here to mile 1.0 we're in a former gravel pit, now an office park. This pit mined deltaic sand and gravel deposited in a glacial lake in the Wills Brook valley, a local tributary of the Musconetcong River. This lake formed as the ice front, retreating from the terminal moraine, dammed the north-draining valley. Spillways for the lake are westward into the main Musconetcong Valley, across the back end of the terminal moraine.

0.9 Proceed straight through light.

1.0 Proceed straight through light.

1.1 Cross I-80.

1.2 Turn right onto Waterloo Valley Road.

1.3 Cross railroad. NJ Transit recently opened passenger service to Hackettstown on this former Erie-Lackawanna branch line. Previously, passenger service stopped at Netcong.

1.6 Road narrows as we leave the Trade Zone complex. We are now entering the back (north) side of the late Wisconsinan terminal moraine, and we will remain on the moraine to mile 7.8. The front of the moraine, marking the farthest advance of late Wisconsinan ice, is about 2 miles to the southwest, on top of Schooleys Mountain.

2.1 Cross railroad again. To right just after railroad there is a meltwater channel that parallels the road. This channel was cut through the moraine by drainage from the lower stage of the lake in the Wills Brook valley.

3.0 View of large sand pit to right, operated by Tilson Aggregates (as of August 2000). This pit, the old pit at stop 1, and the Saxton Falls Sand and Gravel Co. pit across the road from stop 2, all mine lacustrine sand from beneath till of the terminal moraine (see fig. VII-2). The till has been removed from the pit

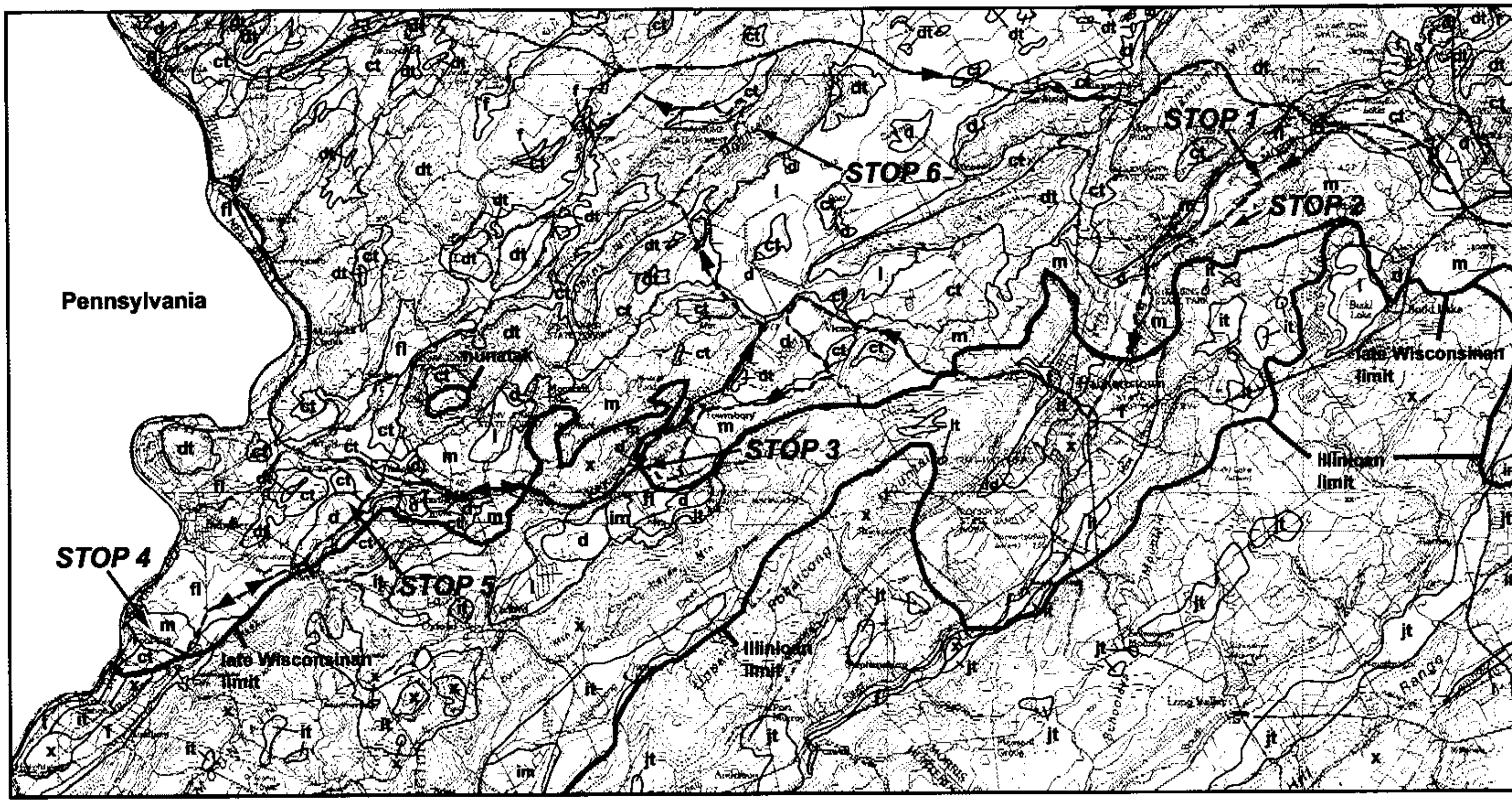


Figure VII-1.--Glacial deposits of the field trip area, showing trip route (dashed line) and stops. Units are: ct=continuous till, t=discontinuous till and rock outcrop, m=till of the terminal moraine, d=deltaic and lacustrine-fan deposits, l=lacustrine clay and silt, f=fluvial deposits, fl=fluvial over lacustrine deposits, it=Illinoian till, im=Illinoian moraine, if=Illinoian fluvial deposits, id=Illinoian delta deposits, jt=pre-Illinoian till, x=nonglacial materials, chiefly colluvium and weathered bedrock.

areas, exposing the underlying sand. The many boulders littering the surface and edges of the pits are relicts of the former till cover. The pits have been dug to their property limits, so there is little face-mining today. Instead, most sand is obtained by drag-lining the lacustrine sand from excavations below the water table in the pit floors.

3.8 Stop 1. Park on right along line of boulders. Walk along path (flagged with surveyor's tape) to assembly point in old pit at A (fig. VII-2). After discussion and examination of outcrops at A, continue on flagged path to large erratic boulders at B and then back to the bus

STOP 1

Leader: Scott Stanford

Setting: Stops 1 and 2 are in the Waterloo Valley segment of the Musconetcong Valley. Waterloo Valley is bordered on the northwest by Allamuchy Mountain and on the southeast by Schooleys Mountain. The terminal moraine runs along the southeast side and floor of the valley, and extends up the east wall of the valley to the top of Schooleys Mountain (fig. VII-2). The moraine loops across the valley on the north side of Hackettstown (fig. VII-1) and then runs westward over Pohatcong Mountain to the Pequest Valley. Exposures in the large sand pits in Waterloo Valley, and logs of wells and borings, show that there is thick lacustrine sand and silt beneath 10 or 20 feet of till on the back end of the terminal moraine on the floor of the valley (fig. VII-2, section AA'). Beneath this sand is a coarser stratified deposit that may be of Illinoian age. Because the lacustrine sand is overlain and deformed by till of the moraine (fig. VII-3), it was laid down in a lake that occupied the valley before ice advanced to the terminal position. The dam and outlet for this lake are uncertain. It is possible that advancing ice crossed Waterloo Valley slightly earlier at its south end, near Hackettstown, forming an ice dam across the valley there. Allamuchy Mountain narrows and lowers northwest of this part of the valley, providing less of an impediment to ice flow than it is to the north. Alternatively, preexisting Illinoian deposits, now buried or eroded, may have formed a dam across the valley.

Striations on Allamuchy Mountain and in Waterloo Valley show that the ice advanced to the south or slightly east of south, flowing out of Kittatinny Valley and across Allamuchy Mountain to the terminal position. This flowpath accounts for the observed mix of gray shale and carbonate rock (from Paleozoic rock in the Kittatinny Valley) and gneiss from Allamuchy Mountain in the drift in Waterloo Valley.

As the ice front retreated from the valley, meltwater streams eroded the moraine dam across the valley at Hackettstown, and any short-lived recessional glacial lakes held in by this dam drained. A glaciofluvial plain (unit Qf on fig. VII-2) was then deposited by meltwater draining from upstream in the Musconetcong Valley. When meltwater drainage ended, the postglacial Musconetcong River incised into the glaciofluvial plain and moraine to form a floodplain.

Materials: At site A there are two outcrops (fig. VII-3) that illustrate ice advance over the lacustrine deposits. The lacustrine deposits (units 2, 3, and 4 on fig. VII-3) are chiefly laminated to thinly bedded very fine-to-fine sand, with occasional medium sand lenses and silt and clay laminae. These beds are typical of suspension deposition from proximal turbidity underflows, most likely in deltaic bottomset beds (Ashley and others, 1985). Hints of ripple cross-bedding in the medium sand lenses indicate an occasional traction current. The till atop the lacustrine deposits (unit 1) has been mostly stripped off during mining, but the surviving fragments are a massive fine-to-medium sand with a few pebbles and cobbles. This till is sandier than the till elsewhere in the moraine and was probably formed by incorporation of the lacustrine sand.

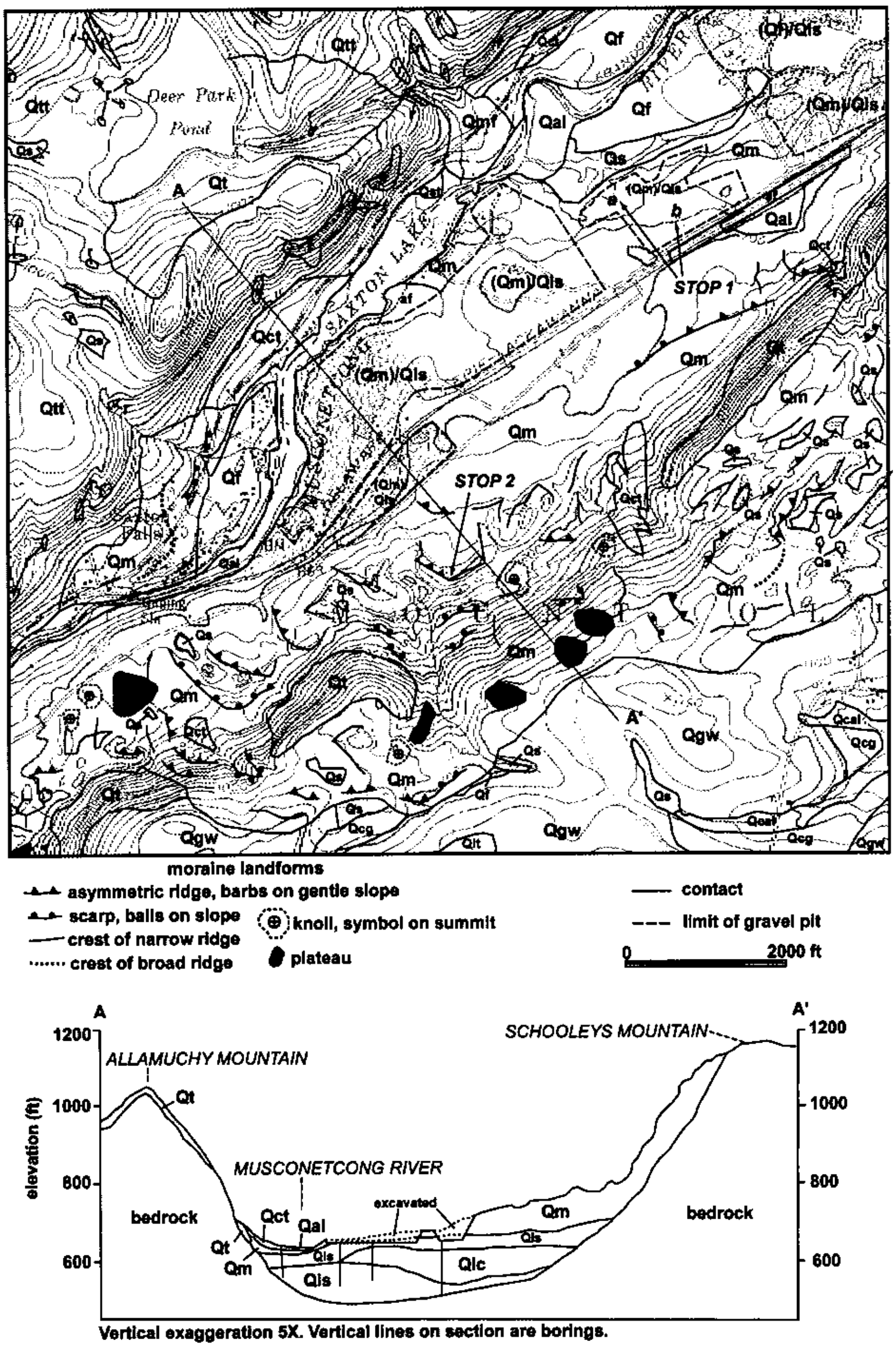
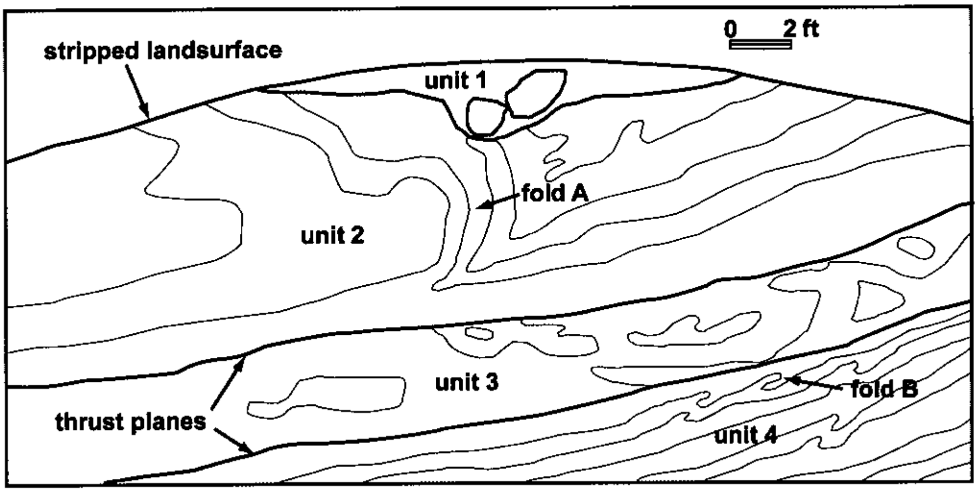


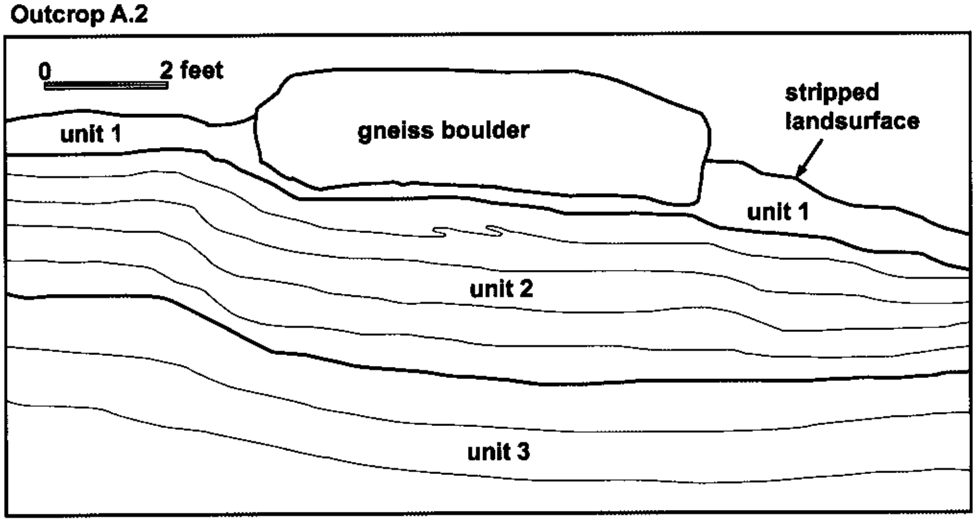
Figure VII-2. Map and section for Waterloo Valley (stops 1 and 2). Map units are: af=fill, Qal=alluvium, Qs=swamp deposits, Qst=postglacial stream terrace deposits, Qct=colluviated till, Qcal=colluvium and alluvium, undivided, Qt=till (>20 feet thick), Qtt=thin till (<20 feet thick) and scattered outcrop, Qm=till of the terminal moraine, Qf=glaciofluvial deposits, Qls=glaciolacustrine sand, Qgw=weathered gneiss, Qcg=gneiss colluvium, Qit=Illinoian till. Subsurface units on section include: Qlc=glaciolacustrine silt, clay, fine sand; Qis=glaciolacustrine sand and gravel of Illinoian age. Units in parentheses have been removed by excavation, exposing units to right of slash, for example, (Qm)/Qls.

Outcrop A.1



unit 1=massive fine-to-medium sand diamicton with cobbles (till)
unit 2=massive to laminated very fine sand with a few silt laminae, with large recumbent fold
unit 3=same as 2, but highly deformed, with recumbent folds extended into boudins
unit 4=laminated fine-to-very fine sand, trace of medium sand, locally with a few small isoclinal recumbent folds
axis of fold A trends N35E, fold B trends about N40E, suggesting ice flow to southeast

Outcrop A.2



unit 1=massive medium sand (till or sheared lacustrine sand)
unit 2=laminated to thinly bedded fine-to-medium sand with a few silt laminae, bent under the boulder and with a few small isoclinal recumbent folds near contact with unit 1
unit 3=laminated to thinly bedded very fine sand and silt, slightly bent under the boulder

Figure VII-3. Sketches of outcrops 1 and 2 at Stop 1A. Heavy lines are contacts of sediment units. Thin lines show bedding forms within units. Scales are approximate. Features at outcrop 1 suggest shear or shove by active ice; those at outcrop 2 suggest passive deposition of boulder on the lake sand.

Discussion: Deformational structures at outcrop A.1 (fig. VII-3) include both large- and small-amplitude recumbent folds, thrust planes, and highly extended recumbent folds that have stretched into boudins. The trend of the fold axes is northeasterly, suggesting ice advance to the southeast, which is in agreement with the striation data and moraine trend. It is likely that the lacustrine sediments were saturated when overrun, and the increase in pore pressure when they were loaded by the overlying ice would have lowered their shear strength and encouraged flow deformation. The recumbent folds may have been produced by the flow deformation, as trapped water was forced out from beneath the ice toward the margin. The thrust planes, on the other hand, suggest a later episode of brittle deformation, perhaps under thicker ice when the sediment was dewatered or frozen.

Outcrop A.2 (fig. VII-3) shows a more passive style of deformation. Emplacement of the large gneiss boulder seems to have bent the underlying lacustrine beds, but there is no evidence for extensive flow or thrusting. A few faint, small recumbent folds beneath the boulder are most likely due to local water escape from underneath the boulder. The absence of load deformation or fluid-escape structures in the underlying sand suggests that the boulder was melted out gradually from the base of the ice, with the weight of the overlying ice preventing upward fluid release.

Return Route: On the return route to the bus, you will climb up a gully in an old pit wall. A large overturned fold, with a northeast-trending axis, was formerly exposed here, and was traceable in the adjoining gullies. This fold also indicated southeasterly ice flow. Later, the return route passes two large erratics on the old pit floor at B. One of the boulders is gneiss from Allamuchy Mountain, the other is a cherty carbonate rock from the Kittatinny Valley. Before mining, these boulders were likely in a similar position as the one exposed at outcrop A.2.

Reference Cited:

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4.1	Cross railroad.
4.5	Saxton Falls Sand and Gravel Co. pit on right.
4.9	Stop 2. Park on left at powerline crossing. Follow jeep trail and footpath (flagged with surveyor's tape) to assembly point at A (fig. VII-4). After discussion, continue on flagged loop route through moraine, passing by the ephemeral pond in the swale at B and then returning to bus.

STOP 2

Leader: Scott Stanford

Setting: At this stop we will follow a walking route through the moraine in order to examine moraine landforms and discuss their genesis. This part of the moraine is at the base of Schooleys Mountain, approximately 0.5 mile back from the front of the moraine at the top of the mountain (fig. VII-2). Although there are no borings here, the lacustrine sand we saw at stop 1 and that is exposed in the Saxton Falls pit across Waterloo Valley Road most likely continues under the area of Stop 2 to the base of Schooleys Mountain (fig. VII-2, section AA'). The

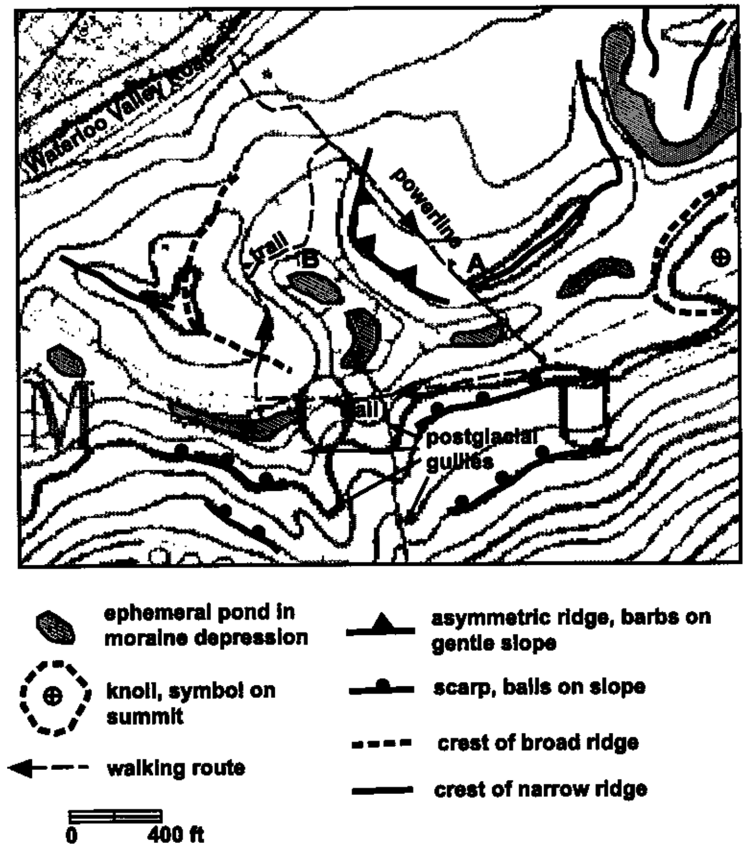


Figure VII-4.--Landform map of stop 2, showing walking route. Contour interval 20 feet.

numerous ephemeral ponds in the moraine swales in this area, unlike those higher on the mountain, drain rapidly and do not have much peat build-up, suggesting that permeable sand underlies the moraine till here.

Materials and Landforms: The terminal moraine is composed of several landform elements, including ridges, scarps, knolls, plateaus, and swales. These features are best identified and mapped from stereo airphotos, and are only hinted at by the topography shown on the quadrangle maps. The landforms shown on figures VII-2 and VII-4 are drawn from 1:12,000 airphotos. Ridges can be subdivided into asymmetric ridges, where one side is steeper than the other, narrow-crested ridges, and broad-crested ridges. Asymmetric ridges, and scarps, tend to be colinear and roughly parallel to the moraine trend. Narrow-crested and broad-crested ridges have more variable orientation, and the narrow-crested ridges often intersect at high angles to define a polygonal network. Knolls are rounded hills that are generally isolated and widely separated from each other. Plateaus are flat-topped hills bounded by distinct scarps. A few have raised rims but most do not. Swales typically are the low areas between parallel ridges, and are crescentic or elongate in shape. A few swales have a circular shape and are depressions in a surface rather than low areas between ridges.

As far as can be determined from excavations and exposures, the material making up all of these landforms is till. The till in Waterloo Valley is a yellowish brown silty sand with many pebbles, cobbles, and boulders. The eroded trailbed at point A on the walking route exposes this till. Much of the matrix material, and most of the gravel and boulder fraction, is derived from gneiss bedrock on Allamuchy Mountain, which is the only gneiss traversed by the ice en route to this location (fig. II-2). The size of the terminal moraine in Waterloo Valley implies that a large volume of rock was eroded from Allamuchy Mountain. The abundance of plucked and abraded rock outcrops, and the absence of colluvial deposits, which are extensive south of the terminal moraine (see mile 9.5-11 of road log), on Allamuchy Mountain attest to the degree of erosion.

Discussion: The origin of some of the landform elements of the moraine is straightforward. Some asymmetric ridges, usually along the front edge of the moraine, are *colluvial ramparts*--aprons of flowtill built up against the ice front. When the ice melts, the ice-contact part of the apron collapses and becomes steep while the debris apron remains as a gentler slope. Scarps along hillsides (for example, those at the bottom of fig. VII-4) are the ice-contact slopes of till embankments deposited as the ice margin retreated downslope. Many of the elongate swales are simply areas of nondeposition between ridge deposits, and do not necessarily mark the position of an ice block as in the familiar kettle-hole model (for a real kettle hole see stop 5). Other landforms may have multiple origins. For example, the sharp-crested ridges that form polygonal networks suggest deposition of flowtill in crevasses or cracks in stagnant ice masses, which then melt to leave the ridges as positive features. Other sharp-crested ridges that are colinear and parallel the moraine trend may be colluvial ramparts that lack asymmetry, or may be push ridges deposited by active, advancing ice. Asymmetric ridges where the gentle slope dips upglacier (like the one on the walking tour) may also be push ridges formed by readvancing ice, possibly reshaping previously deposited till. Rounded, gently sloped landforms, like some of the broad-crested ridges and knolls, may have been shaped subglacially. Streamlining beneath moving ice accounts for their smoothed slopes. The plateaus have been interpreted elsewhere as ice-walled lakes or ponds. However, those in New Jersey are not known to contain lake sediment and instead seem to consist of till. Their genesis is obscure. Perhaps the planar top indicates a partly subglacial origin for these landforms.

While definitive origins for each landform may not be decipherable, it seems clear that the moraine includes a mosaic of active and stagnant ice features. Radiocarbon dates (fig. II-8 and table II-1), and the large volume of the moraine (at least an order of magnitude larger than any of the recessional moraines in the northeastern U. S.), indicate that deposition of the moraine may have taken as much as 2,000 years. Over this length of time, the ice front is sure to have alternately retreated and readvanced repeatedly within the moraine belt. With retreat, debris-covered ice masses would have become isolated and separated from the main glacier, eventually melting to form stagnant-ice topography. With advance, earlier deposits, including stagnant ice blocks, would have been overrun and reshaped into active ice-front and subglacial forms.

Consider these alternatives as you traverse the landforms and return to the bus. Refer to chapter III for further discussion of moraine genesis.

6.3 Cross Musconetcong River. From here southward through Stephens State Park to Hackettstown the river flows in a narrow postglacial valley notched about 60 feet into the terminal moraine. Turn left at stop sign onto county route 604 south.

7.6 Begin descent down front of terminal moraine.

7.8 Cross from front edge of terminal moraine onto glaciofluvial plain. This glaciofluvial deposit was laid down while ice stood at the terminal moraine, and extends (with some interruption due to postglacial erosion) down the Musconetcong Valley to merge with the late Wisconsinan glaciofluvial deposit in the Delaware Valley at Riegelsville.

8.6 Rise from plain onto a carbonate-rock upland.

8.7 Turn right at light onto US 46 west.

8.9 Illinoian till veneers the rock surface in this swale.

9.3 Cross railroad and pass onto a small late Wisconsinan glaciofluvial plain along Hatchery Brook. This plain, which is not in contact with an ice margin, was fed by meltwater draining from the terminal moraine on Pohatcong Mountain to the northwest.

9.5 Rise from plain onto apron of gneiss colluvium along base of Pohatcong Mountain. These aprons are common along the bases of steep slopes south of the terminal moraine. Sedimentary features of the colluvium suggest that creep and solifluction were the primary depositional processes. These processes, in turn, are most active when permafrost is present, so most of the colluvium is of periglacial origin. Remain on colluvium until mile 11.0.

11.0 View to left of marsh dammed to the east by colluvium, and to the west by the terminal moraine. Cross from colluvium and reenter terminal moraine.

12.8 After descending the west side of Pohatcong Mountain, in a topographically subdued part of the moraine, cross onto a flat delta surface in the town of Vienna. This delta was deposited in the earliest, highest stage of Lake Pequest, which was dammed by the terminal moraine where it crosses the Pequest Valley at Townsbury.

13.2 Cross Pequest River, return to delta.

13.7 Turn left off U.S. Route 46 West onto Cemetery Road. The low area to the west (right side) is the southeastward edge of the floor of Lake Pequest. The gravelly, higher ground to the east (left side), is an ice-contact delta laid down in Lake Pequest (stage 1). The road here follows the ice contact slope of the delta.

13.9 At bend in road climb onto ice-contact slope of delta.

14.3 Pass entrance to Pequest Union Cemetery. The flat area on either side of the road is a delta plain, which lies at an elevation of 585 feet (a.s.l.). In places, the delta plain contains kettles, topographic depressions formed by melting buried blocks of stagnant ice. Deltas consist of planar sets of topset beds that consist of coarse gravel and sand overlying foreset beds of fine gravel and sand. Near the meltwater feeder stream, foreset beds are generally steeply inclined (25° to 35°) and consist of thick to thin, rhythmically-bedded fine gravel and sand. Farther out in the lake basin these sediments grade into less steeply dipping foreset beds of graded, ripple cross-laminated and parallel-laminated sand and fine gravel with minor silt drapes. These in turn grade into gently dipping bottomset beds of ripple cross-laminated, parallel-laminated sand and silt with clay drapes.

14.4 Good view of delta plain and kettles looking southward down the Pequest Valley (right side).

14.5 Descend delta slope and cross Pequest River. Glacial lake varves are exposed along the river banks beneath thin alluvium. Glacial varves consist of stacked annual layers that consist of a lower "summer" layer of chiefly silt that grades upward into a thinner "winter" layer of very fine silt and clay. Most of these materials were deposited from suspension, although the summer layer may contain sand and silt carried by density currents. Each summer and winter couplet represents one year. The broad lowland east of the delta is underlain by lake-bottom deposits and till. Till in the valley here suggests that part of the valley floor was covered by ice during the deposition of the Pequest Union Cemetery delta.

14.8 Climb till-covered hillslope behind the Terminal Moraine.

15.2 Turn right onto Townsbury Road (heading south down the Pequest Valley). We have crossed onto the inner margin of the Townsbury segment of the Terminal Moraine. Note the transition to hummocky (knob-and-kettle) topography, the occurrence of large stone rows, and the many large boulders used for landscaping in nearby yards. The upland to the east (left) side of the road is County House Mountain. To the west (right side) is Danville Mountain and Mount Mohepinoke, which make up part of the Jenny Jump Mountain thrust sheet. The Pequest Valley narrows considerably here (< 4000 feet) from its much broader width above Great Meadows (> 8000 feet).

15.6 Pass entrance to small housing development on west (right) side of road. The flat area beneath the homes is a small ice-contact delta that was deposited in lake Pequest. It has the same elevation as the Pequest Union Cemetery delta.

16.1 Drive beneath power lines and climb onto a small morainal knoll. Knob-and-kettle topography is well formed in this area of the moraine.

16.2 Cross small bridge over postglacial channel cut in moraine.

17.2 The steep scarp on the west (right) side of the road was chiefly cut by meltwater draining from Lake Pequest.

17.3 Stop sign. Intersection of Townsbury Road with Pequest Road. Turn left onto Pequest Road and leave main part of morainal belt. The broad lowland in front of the Terminal Moraine here is underlain by meltwater-terrace deposits and remnants of the moraine. Meltwater-terrace deposits were chiefly formed on eroded parts of the moraine. In many places they consist of a boulder lag. Erosion of the moraine may have occurred during an outburst of Lake Pequest when flood waters, related to a breach in the moraine dam (Ridge, 1983), scoured the valley floor.

18.0 Pass over a small cross-valley ridge (remnant of the Terminal Moraine).

18.3 Intersection of Pequest Road and Janes Chapel Road. Bear right and stay on Pequest Road. The small structure to the west (right) side of the road is a well house for one of the Hatchery's supply wells. Lower flat surfaces in the valley are meltwater terraces, and the low hills to the east (left side) are remnants of ice-contact deltas laid down in Lake Oxford.

18.5 Right turn into rear entrance of Pequest Fish Hatchery. The entrance is near the late Wisconsinan border in the Pequest Valley, which is marked by small ice-contact deltas on either side of the valley. Depart buses at nearby parking area on left and head over to picnic area across from parking lot. Lunch time!

STOP 3

Lunch and a brief overview of the local geology and hatchery operations.
Summary by Richard Dalton
New Jersey Geological Survey

The selection of the Pequest site for the proposed fish hatchery was based on field geology, pump tests, and water recorder data. A number of other localities in northern New Jersey were studied while looking for a suitable hatchery site.

A modern fish hatchery that will produce trout by the most efficient method and at the least cost, requires about seven thousand gallons of water per minute. Geologic field work, research and testing of the Pequest area by the New Jersey Geological Survey began in the Summer of 1971. This part of the Pequest Valley was one of the few areas in central or northern New Jersey capable of yielding 7,000 gallons per minute. Since the basic ingredient of a fish hatchery is sufficient water of a suitable quality and temperature, it is critical that this supply is sustainable during droughts and that it can be protected from pollution.

The Pequest Valley near the hatchery is underlain by a complex assemblage of Illinoian and late Wisconsinan glacial deposits that rest on dolostone (figs. VII-1 and VII-5). Illinoian deposits include till and recessional moraine. These materials in places lie on weathered dolostone (saprolite). All these surficial deposits have low permeability. Late Wisconsinan deposits consist of basal till, and till that forms the Terminal Moraine. These materials also have low permeability, although the mix of stratified materials in the moraine may increase permeability locally. Meltwater deposits include deltaic sand and gravel, and lake-bottom deposits of laminated silt and clay that were laid down in Lake Oxford (see Chapter III for a detailed description of the lake). The sand and gravel, delta topset beds and foreset beds, are highly permeable whereas the lake-bottom deposits have a very low permeability. In the lower parts of the valley, thin, gravelly meltwater-terrace deposits and alluvium cap the deltaic materials.

This ground water system in the Pequest Valley receives water from the surrounding uplands, which are underlain by fractured gneiss and granite, and also the area up valley by transmission through very permeable glacial sand and gravel, and limestone. Based on projected ground water withdrawals, and pump test data and water recorder information gathered over a two and one-half year period, a minimum spacing between wells was calculated. Because of these spacing requirements, it was necessary to acquire additional lands for the Pequest Fish Hatchery. The reasons that the additional lands were acquired are given below.

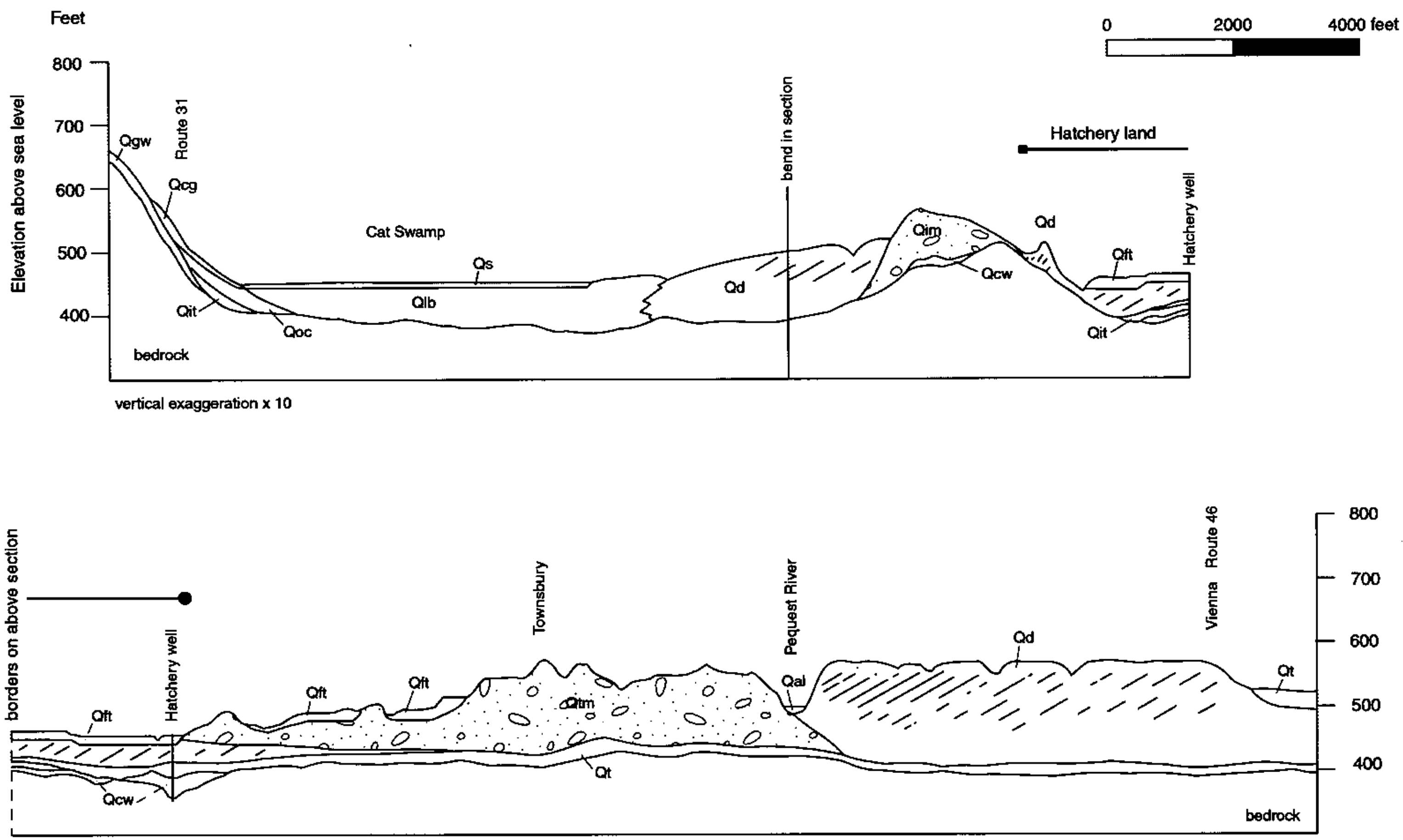


Figure VII-5. Longitudinal section up the Pequest Valley showing surficial geologic formations and stratigraphic framework of valley-fill materials in the vicinity of the Pequest Fish Hatchery, Stop 3. List of units: Qal - alluvium, Qs - swamp deposits, Qd - ice-contact delta, Qlb - lake bottom deposits, Qft - meltwater-terrace deposit, Qtm - Terminal Moraine, Qgw - weathered gneiss and granite, Qcw - weathered carbonate rock, Qcg - colluvium derived from weathered gneiss and granite, Qoc - older weathered colluvium, Qit - Illinoian till, Qim - Illinoian recessional moraine. Figure modified from Witte and Stanford (1995).

1. Based on data obtained during a 10-day pump test, determinations on aquifer characteristics were made. The calculations show that a minimum spacing between the wells be at least 1000 feet, with the bulk of the well field extending diagonally away from the river up the valley. This minimum well spacing is needed to prevent infiltration from the river during times of drought. A spacing of less than 1000 feet will cause a substantial lowering of the water table near the river below the level of the river and thus inducing recharge from the river. Any contamination, which may be present in the river, would then be drawn into the aquifer.

2. The permeability of the valley fill is so great (more than 1000 ft. per day in this area) that any development near the well field, either industrial or residential, would result in rapid degradation of the aquifer. Controlling the amount and types of agricultural chemicals being deposited on the surface near the well field is also important since they would be carried rapidly into the system by downward percolating waters. Because of the rapid subsurface transmission, pollution could, if the surrounding area were not owned by the State, spread to the hatchery within a matter of hours.

In the proposed hatchery design, practically one hundred percent of the water used in the fish rearing process will be returned to the water system, thus maintaining a nearly equal natural system. Encroachment on this natural geohydrologic system by any facility other than a State fish hatchery will surely result in water reduction and probable degradation.

Return to buses after a 45 minute lunch break.

19.1 Entrance road to Hatchery (cross over Pequest River).

19.2 Turn left onto U.S. Route 46 West from the main entrance to Hatchery. Low terrace near the highway is a meltwater terrace. Mount Mohepinoke forms the large upland to the north west (right side of the highway). At this point we are outside the late Wisconsinan border. Most of the terraces in the valley here were formed by outflowing waters from Lake Pequest.

20.5 Intersection of Pequest Furnace Road and U.S. Route 46 (left side of the highway). Enter east side of Pequest Gap.

21.0 Intersection with Free Union Road (right side of the highway). Cross back over the late Wisconsinan border. The higher ground to the north (right side) is the Mountain Lake segment of the Terminal Moraine. Pequest River lies to the south (left side). When ice advanced to the late Wisconsinan border it blocked drainage on the west side of Pequest Gap forming Lake Oxford. During retreat to a position marked by the Terminal Moraine, a lower outlet was uncovered through the gap. Lake drainage from both Lakes Oxford and Pequest cut a channel through the Mountain Lake and Bridgeville segments of the Terminal Moraine.

21.7 Cross Mountain Lake Brook.

21.8 Stop light, intersection of U.S. Route 46 and State Route 31. Continue straight on U.S. Route 46 West. The low hills to the north (right side) make up part of the Mountain Lake segment of the Terminal Moraine. The valley floor here is covered by meltwater-terrace and stream-terrace deposits that form a sequence of terraces that extend far down the valley.

22.4 Johnny's Hotdogs on the left. The Terminal Moraine in this area consists of the Mountain Lake and Bridgeville segments, which were formed by sublobes of the Kittatinny Valley ice lobe flowing around and over Jenny Jump Mountain.

22.5 Dolomite outcrops on north (right) side of U.S. Route 46.

22.7 More dolomite outcrops on north (right) side of U.S. Route 46.

23.1 Stop light. Intersection of U.S. Route 46 with County Route 519. Turn left onto County Route 519 South.

23.3 Cross over Pequest River.

23.7 Pass entrance to Unangst delta on west (right) side of County Route 519. Site of stop five.

23.8 White Township Garage. Southward dipping foreset beds of the Pophandusing delta are exposed in small excavation behind the garage. County Route 519 follows the floor of a meltwater channel cut down in the Pophandusing delta. The channel served as an outlet for a small proglacial lake that formed behind the delta and the Bridgeville segment of the Terminal Moraine and the Delaware Valley sublobe. The late Wisconsinan border extends about 1000 to 1500 feet up the slope to the southwest (left side) of County Route 519.

24.8 Stop light. Intersection of County Route 623 (Brass Castle Road) and County Route 519. Continue straight on Rt. 519 South. The broad valley floor is underlain by a thick sequence of glacial outwash. Well records list clay and silt lying beneath sand and gravel. The fines are interpreted as glacial lake-bottom deposits, laid down in a proglacial lake that formed between the Foul Rift segment of the Terminal Moraine and the retreating glacier. Several levels of terraces cut in this surface are probably related to erosion of the Foul Rift moraine, and lowering of local base level.

26.4 Stop sign. Intersection of County Routes 620 (Belvidere Road) and 519. The Martins Creek power plant lies straight ahead. It is an oil fired facility (no nukes here!). Turn left and continue on Route 519 South.

26.5 Pass entrance to River Edge Park. Coarse boulder-cobble gravel was exposed during the construction of the town houses. These materials are part of an outwash head that is coeval or slightly younger than the Foul Rift moraine.

26.6 Kettled outwash plain on east (left) side of the road. The New Jersey Highlands form the massive upland to the east.

26.7 Pass entrance to Warren County Garage. The low hills behind garage make up part of the Foul Rift moraine.

26.8 Pass entrance to Warren County offices on west (right) side of the road. The office complex sits on the moraine. The highway here crosses a low rise that forms the eastern edge of the moraine. The low area to the east (right) is underlain by outwash.

27.1 Turn right onto Foul Rift Road heading west. We will cross over three outwash terraces that increase in height toward the west. The low hills to the north (right) make up part of the moraine.

27.4 The low ridge in front of the moraine is a till-covered bedrock (dolomite) ridge. The late Wisconsinan border extends about 2500 feet down the Delaware Valley (left) from the moraine.

27.5 Cross onto the Foul Rift moraine. Ridge and trough topography is well formed on the north (right) side of road.

27.8 Turn right into entrance to the Foul Rift pit (owned and operated by Harmony Sand & Gravel). Park in area across from scale house. We'll assemble in the parking area and quickly march to the lower section of the pit by way of the road past the scale house. To keep on schedule, we have about an hour for this stop. This is a working pit that I have had access to for the last fifteen years. Harmony Sand and Gravel has graciously consented to the use of this pit as one of our field stops. For safety considerations please avoid the pit walls and equipment. Our route down into the pit will take us by several stock piles of variously-sized aggregate. Harmony Sand and Gravel processes glacial outwash and some materials that make up the end moraine for a variety of construction and landscaping applications (select fill and stone, roofing aggregate, landscaping stone, concrete and masonry sand, road base, rip rap, and building stone). The huge stock pile of cobbles on our right, as we descend into the pit, has been growing for at least fifteen years. Due to the large size of the stones they are of limited economic value. Future plans involve crushing the cobbles to a more useful size. Pass the cobble stock pile (left side of road) are several pieces of equipment used to wash and sort the variously-sized materials mined from this pit.

STOP 4

Foul Rift Segment of the Terminal Moraine
Leader: Ron Witte

Setting: The Foul Rift pit lies in the Delaware Valley about two miles south of the village of Belvidere, New Jersey (fig. VII-1). The valley is underlain by Cambro-Ordovician dolomite and limestone of the Kittatinny Super Group and the Jacksonburg Limestone, an Ordovician muddy limestone that is quarried and processed into Portland Cement. These formations strike southwest, a direction also followed by the Delaware River in this area. Locally, the Delaware Valley consists of two smaller subvalleys (Buckhorn Creek and Delaware River), which are separated by a ridge of dolostone that extends about three miles down valley to Hutchinson, New Jersey. The New Jersey Highlands, held up by Proterozoic metasediments and intrusive rocks, border the valley on the east. The lower lying hills to the west are held up by the Martinsburg Formation, an Ordovician interlayered sequence of slate and graywacke. In the past, slate has been extensively quarried for use as roofing slate. This industry has seen a sharp decline in the last 50 years due to the use of manufactured materials for roofing. The pit is named after a nearby section of class two rapids on the Delaware River formed over dolomite ledge and glacial boulders. Well records (Witte and Stanford, 1995) show that the pre-late Wisconsinan course of the Delaware River lies eastward beneath Buck Horn Creek valley. The present course of the river through Foul Rift and downstream to Hutchinson, New Jersey was cut by meltwater during the last glaciation.

Glacial deposits of Illinoian and late Wisconsinan age are found in the Foul Rift area (figs. VII-1 and VII-6). Illinoian till caps the rock ridges south of the late Wisconsinan border and Illinoian outwash forms a terrace above late Wisconsinan outwash down valley near Brainards, New Jersey. Illinoian outwash and colluvium of Sangamon and Wisconsinan age have also been found beneath late Wisconsinan outwash by Ridge and others (1986) and the author (unpublished data, New Jersey Geological Survey, Trenton, New Jersey) downvalley from the late Wisconsinan border. South of the Foul Rift moraine, late Wisconsinan outwash forms a large valley train that extends far downstream to Trenton, New Jersey. Just east of the moraine the valley train is incised by several meltwater terrace deposits. Meltwater that carved these terraces emanated from ice retreat positions

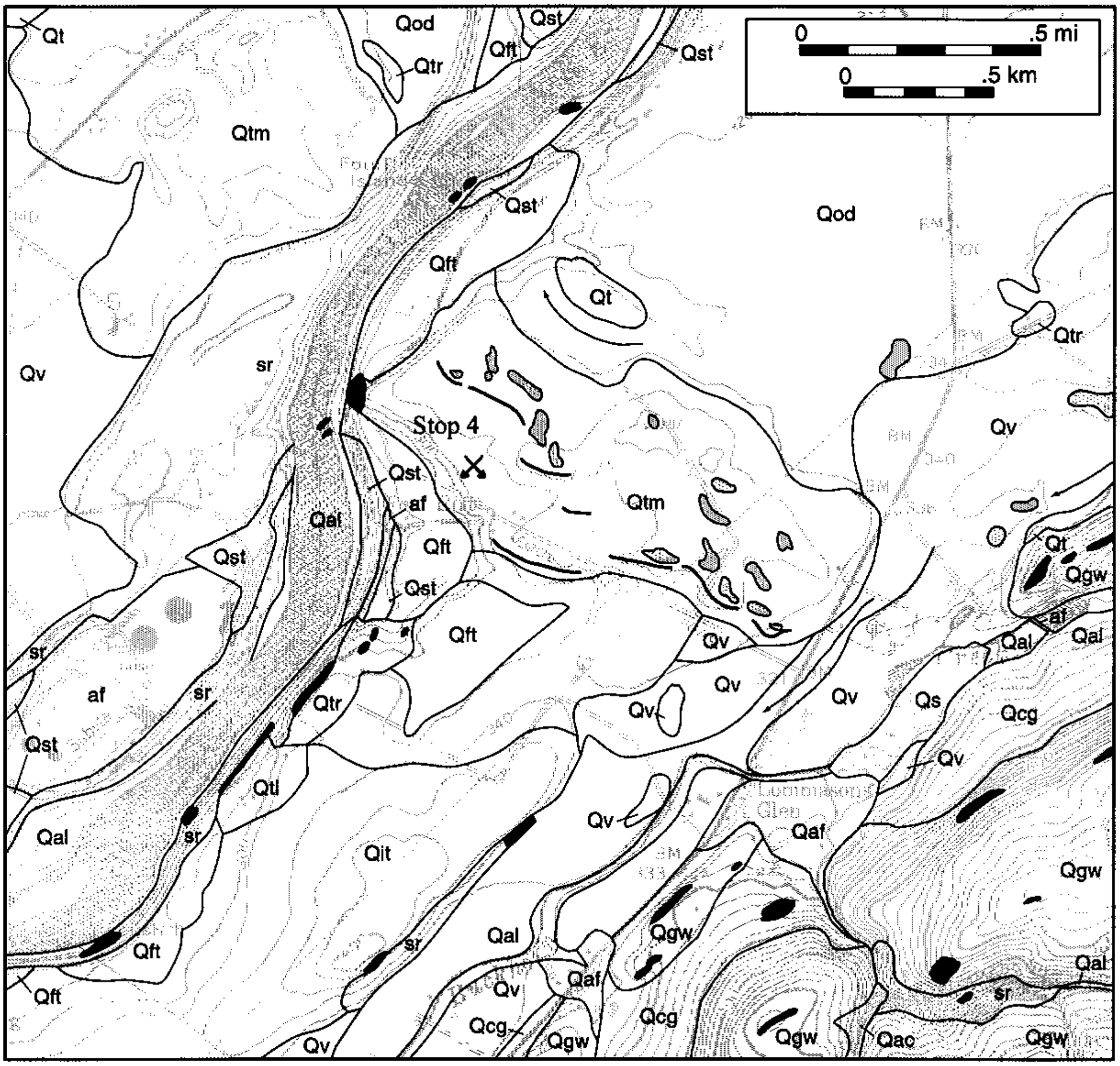


Figure VII-6. Surficial geologic map of part of the Belvidere, NJ-Pa, 7 1/2 minute topographic quadrangle near Foul Rift and location of stop 4. List of map units: af - artificial fill, Qal - alluvium, Qaf - alluvial fan, Qst - stream-terrace deposit, Qs - swamp deposit, Qac - alluvium and colluvium undifferentiated, Qtm - Terminal Moraine, Qt - thick till, Qtr - thin till, Qv - valley-train deposit, Qod - valley-fill delta, Qft - meltwater-terrace deposit, Qit - till of Illinoian age, Qtl - till-stone lag, Qcg - gneissic and granitic colluvium, Qgw - weathered gneiss and granite, and sr - shallow rock. Areas shaded black represent bedrock outcrops, and areas shaded light gray are kettles. Heavy dark lines on the moraine denote the crests of moraine-parallel ridges and thinner lines with an arrow denote meltwater channels. Data modified from Witte and Ridge (in preparation), Surficial geologic map of the Belvidere Quadrangle, NJ - PA.

farther up valley (see fig. III-19 for a description of how these terraces may have formed). The Foul Rift moraine forms a large cross-valley ridge that exhibits ridge-and-trough and knob-and-kettle topography (fig. VII-6). In New Jersey, its western edge forms an erosional scarp against the Delaware River, while its eastern edge terminates against outwash. Behind the moraine outwash fills in the broad valley south of Belvidere. This material lies below the moraine and nearby valley-train terrace. Well records (Witte and Stanford, 1995) show fine sand and silt at depth. Apparently, the moraine dammed the valley at Foul Rift and a lake formed in front of the retreating Delaware Valley sublobe. Over time deltaic outwash materials filled the lake basin from valley wall to valley wall. Lower meltwater-terraces deposits cut in the valley-fill delta mark erosion of the moraine dam and lowering of local base level in the valley. Eventually drainage shifted to the present course of the Delaware River.

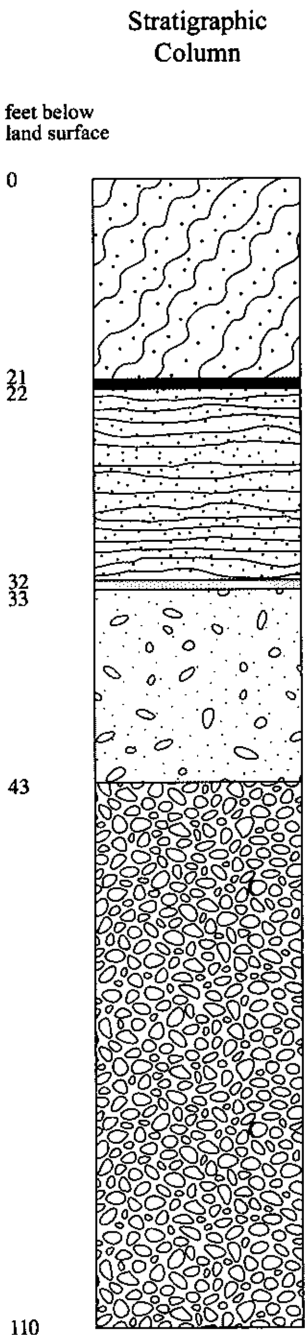
Discussion: The Foul Rift pit provides an exceptional opportunity to study in three dimensions, glacial valley fill laid down at the front of the Kittatinny Valley ice lobe, and view a cross section of an end moraine. Figure VII-7 represents a measured section that summarizes the Foul Rift stratigraphy. Unfortunately, this section is now covered by a very large stock pile of cobbles and small boulders. Because this pit is active, most of the outcrops viewed earlier by the author are no longer available for inspection. However, the overall stratigraphy of the pit and character of materials exposed here have remained fairly consistent. Preparation for this field trip and the observation of several new, but shot-lived exposures, have revealed a few surprises.

The lower gravel and sand (figs. VII-7 and VII-8, inset) is glacial outwash laid down in front of the advancing ice sheet. This unit makes up more than half the stratigraphic section at the Foul Rift pit. Based on the distribution of nearby bedrock outcrops and a reconstruction of the buried rock topography (Witte and Stanford, 1995) bedrock beneath the pit floor is about 200 feet above sea level, rising upward to the east. Based on this estimate there may be an additional 50 to 75 feet of material beneath the lowest part of the pit floor. These stratified materials consist chiefly of matrix-supported planar to cross-stratified cobble-pebble gravel, pebble gravel, pebbly sand, and minor lenses of sand. The provenance of the outwash has a decidedly Delaware Valley lithology. Clasts consist chiefly of dolostone, slate, graywacke, quartzite, with secondary amounts of red sandstone. Gneiss and granite (Highlands source) account for less than 5 percent of the gravel fraction.

These materials were laid down by anastomosing sets of meltwater streams that formed a braided pattern of channels and bars across the valley floor. Most of the gravelly beds are bars, while most of the sandy lenses are channel-fill deposits. Individual beds may be as much as five feet thick, although most are less than two feet thick. Both normal and reverse grading may be observed and grain size changes rapidly in both vertical and horizontal directions. This shows that these materials were deposited under highly fluctuating water discharges, the result of daily and seasonal meltwater production. In a nearby pit, located about one mile downstream in the Buckhorn Creek Valley, boulders, some as large as 3 feet in diameter, form bouldery beds. These features were probably deposited during a meltwater mega-flood related to an outburst of subglacially trapped meltwater. In places the stratified materials are cemented with calcium carbonate. The location of the cemented gravel probably represents the former water table where calcium carbonate was precipitated during cycles of wetting and drying. The cementing agent was probably derived from weathered clasts of carbonate rock.

Lying above the proglacial outwash is a compact, fissile, sandy-silty till that contains many striated and rounded clasts (as much as 15 percent by volume). Lithology of the clast fraction is similar to that of the underlying outwash. In places the till contains thin beds, small lenses, and clots of sand, pebbly sand, and pebbly gravel. These intra-unit materials exhibit horizontal to subhorizontal attitudes, typically have pinch and swell boundaries, and have the overall appearance of having been sheared. The lower till contact is typically abrupt and there is very little mixing across boundaries, other than a few clasts that straddle the contact. Elongated clasts show a preferred long axis orientation downvalley (fig. VII-7). This material appears to be a basal till laid down at the base of the ice sheet when it advanced to its most southern position about 2500 feet downvalley from the pit. A

Section FR-4 - Foul Rift sand and gravel pit



Description of Materials

Upper till - Noncompact, stony, silty-sandy diamicton that contains by volume 15 to 25 percent, chiefly subrounded to rounded cobbles, pebbles and boulders. Contains intrabeds of sheared, stratified pebbly sand, gravel, and laminated silt and clay. Boulders in places form faint layers. Dolomite and sandstone clasts typically exhibit striations. These materials form the surface deposits of the Foul Rift moraine. Origin of materials is polygenetic ranging from an ablationary cover to ice thrusted substrate.

Promorainal outwash - Sheared beds of pebbly sand and sand. Deformation is much more pronounced in the upper part of this unit and was caused by overridding glacial ice.

Promorainal outwash - Mix of cobble-pebble, and pebble gravel, pebbly sand, and laminated silt and clay. Texture and primary sedimentary structures show that these materials were laid down by meltwater streams in both fluvial and lacustrine settings.

Lacustrine sediment - Laminated clay and silt. Deposit largely derived from the settling of fines in a glacial lake or pond. Clay - silt couplets may represent annual layers.

Basal till - Compact, sandy-silty diamicton that contains by volume 10 to 15 percent, chiefly subrounded to rounded pebbles, cobbles, and a few boulders. Contains intrabeds and lenses of sheared sand, pebbly sand, and pebble gravel. Dolostone and sandstone clasts exhibit striae and elongated clasts have a preferred orientation of northeast to southwest (see inset). Deposited at glacier's base. Largely derived from glacially eroded outwash.

Proglacial outwash - Planar to slightly curved, non-parallel tabular to lenticular beds of cobble-pebble, pebble gravel and pebbly sand, and minor lenticular beds of planar to cross-stratified sand. In places cemented by calcium carbonate. Gravel beds are typically less than 3 feet thick, generally rest on erosional contacts, have coarse clasts at their base, and exhibit normal grading. Imbrication of clasts is minor. Sand beds are generally less than 1.5 feet thick, and less than 10 feet wide. Their base conformably overlies coarser gravel and sand, where as the top of the beds have been truncated by erosion. Gravelly beds are channel bars and the sand beds are channel fills. These deposits were laid down by anastomosing sets of meltwater streams.

clast fabric - basal till

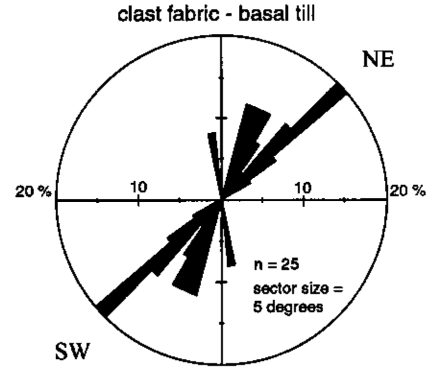


Figure VII-7. Stratigraphy of late Wisconsinan glacial drift measured in the Delaware Valley near Foul Rift, New Jersey. Inset figure is a frequency rose diagram representing the azimuth of elongated pebbles (three to five inches in length) in the basal till.

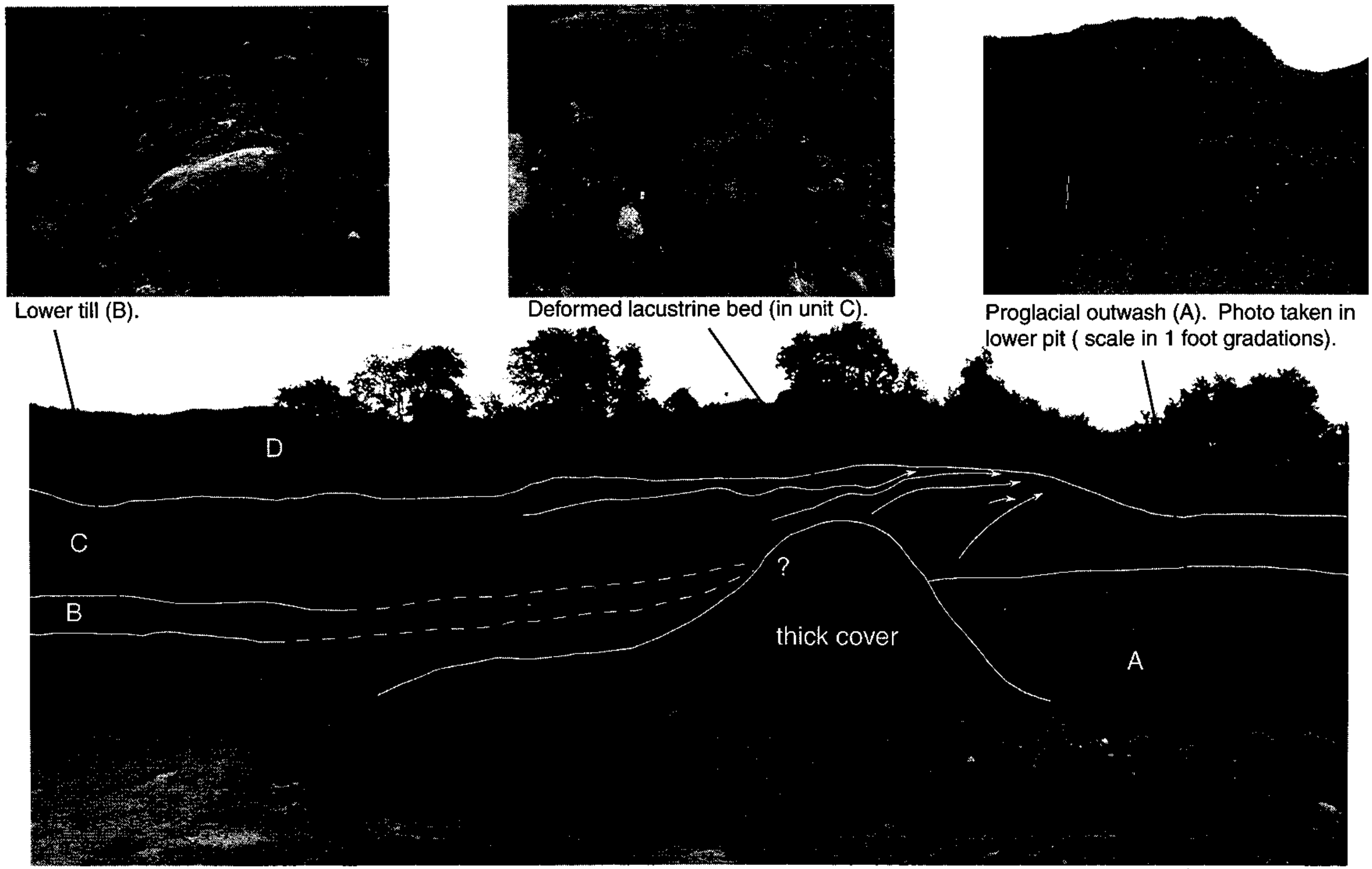


Figure VII-8. Composite section of the east wall, Foul Rift pit. Units: A - proglacial outwash, B - lower till (basal), C - deformed outwash, arrows denote thrusts, and D - upper till (Foul Rift moraine). Thrusts typically marked by deformed beds of silt and clay.

large part of the till is reworked glacial outwash. The preservation of some primary bedding structures within these intra-unit beds suggest that some parts of the outwash were frozen before their incorporation within the glacier's sole.

Based on the location of the lower till and that it was deposited during glacial advance, it may be part of the same till sheet that covers the rock ridge south of the Foul Rift moraine. In places a thin layer of laminated silt and clay caps the till. This unit has been observed elsewhere in the pit. Previous exposures did show that this material extended several hundreds of feet southward. The silt and clay bed might be lacustrine, although its depositional setting is unclear. Its location atop the basal till suggests that the glacier had retreated north of the pit location at the time the fine-grained material was deposited. Possibly residual ice downvalley may have temporarily dammed a lake in front of the ice sheet. These materials may have also been deposited in shallow depressions formed on the surface of the till sheet by glacial scour.

Lying above the till is a complex section of stratified gravel, sand, and silt. This unit has been the most difficult to decipher in terms of its history because its composition, bedding, and geometry vary throughout the pit. Previous exposures showed that materials ranged from cobble-pebble gravel to clayey silt. Bedding contacts typically exhibit pinch-and-swell traces across the outcrop face, and are sharply truncated in places. Texture and sedimentary structures show that these materials were laid down by meltwater streams in both fluvial and lacustrine settings. However, the complex geometric relationships between the various layers and lenses of material cannot be explained as a product of deposition, but a product of postdepositional deformation. Given that these materials were laid down at the margin of an ice lobe and that active ice was present in the Delaware Valley, as shown by the lower till, deformation was probably caused by ice shove during a readvance of the Delaware Valley sublobe.

New exposures (fig. VII-8, located northeast of section FR-4) opened this summer show large scale recumbent folds and imbricate thrusts that further support the contention that this material has been ice shoved and probably overridden by ice. Near the upper center of the figure and below a moraine parallel ridge, there are several stacked ramp-like structures that decrease in attitude going upwards. Beds of darker-colored material highlight these features. They consist of deformed clayey-silt (fig. VII-8, inset) similar to the fine-grained materials previously described. In other places the dark-colored beds are till. Most of the deformation is best preserved in the finer-grained materials rather than the coarser gravels. Apparently the gravelly material was largely deformed by intergranular rotation and sliding (similar to shoving a pile of marbles) where as the finer material, because of its higher moisture content, and competence, deformed more ductilely.

The ramp-like structures may be a sequence of stacked thrusts that consist of ice-shoved outwash and glacial pond sediment. The stacking may represent several advance and retreat cycles, or the thrusts may have developed during a solitary readvance. Based on the amount of deformation observed in section FR-4 (fig. VII-7) it appears that deformation attenuates down valley with the highest degree of deformation occurring beneath a moraine-parallel ridge.

The upper unit (figs. VII-7 and VII-8) consists of a poorly compact, stony silty-sandy till containing lenses and layers of gravel and sand. Some of these materials may be debris flows or flowtill, others are glaciotectonized outwash. Stoniness varies widely throughout the till and dolostone and sandstone clasts are typically striated. In places boulders form weak layers. This upper till unit makes up the Foul Rift moraine. In most places this material looks to have been derived from an ablationary cover at the glacier's terminus. New exposures cut into the northern pit wall suggest that the moraine may in part consist of basal till. To more effectively mine the underlying sand and gravel, the upper till has been stripped by cutting a bench into the northern pit wall. This process has unfortunately covered excellent exposures of basal till and deformed outwash, but it has revealed a character of the upper till never seen before. The character of the moraianal till exposed here (fig. VII-9) is much different than it is elsewhere in the pit (fig. VII-8). It has a more massive structure, and lower stone content. Its matrix is a compact silty sand, and it contains a mix of rounded to subangular stones. Elongated clasts also have a moderately strong down valley fabric (fig. VII-9) and clasts dip in an up valley direction. Based on its character this material appears to be a basal till, although it forms the surface deposit that makes up the moraine.

Interpretations about the moraine's genesis have changed considerably over the years. Most of the earlier workers (Lewis, 1884; Salisbury, 1902) suggested that the cross valley ridge at Foul Rift was part of the Terminal Moraine. Ward (1938) proposed that the moraine was a recessional kame complex formed behind the terminal moraine, which he placed farther downvalley. Ridge (1983) suggested that the moraine was a frontal kame complex (coeval with the Terminal Moraine), largely consisting of stratified gravel and sand laid down among stagnant ice at the margin of the Delaware Valley sublobe. This interpretation was revised by Ridge (1985) upon seeing new exposures. He proposed that the Foul Rift moraine was made up of several push moraines, largely derived from the underlying outwash. The new interpretation emphasized the role of active ice at the glacier's margin.

The recent exposures along the east wall of the pit provide additional insights concerning the formation of an end moraine. The outcrop face shown in Figure VII-8 runs nearly perpendicular to the Foul Rift moraine and it bisects a moraine parallel ridge. This view of the moraine and underlying materials provides strong evidence that active ice associated with one or several readvances formed large parts of the end moraine. Deformation seen beneath the moraine parallel ridge suggests that this feature is the result of ice shove. The ridge has formed where thrusted outwash had formed a submorainal ridge. The till that makes up the moraine is superincumbent on this ridge, possibly as an ablationary mantle of flow till, debris-flow deposits, and blocks of overridden substrate. The basal till (fig. VII-9) that lies along the north rim of the pit (left of fig. VII-8) reflects a change in facies, going from a supraglacial setting to a subglacial setting.

The submorainal ridge may have a deeper origin. The basal till measured in section FR-4 appears to have been a continuous sheet based on older exposures seen by the author. In Figure VII-8 (inset) the lower till crops out on the lower left side of photograph. From here it was seen to rise as much as 15 feet and eventually terminate near a position well below the imbricate thrusts. It may lie beneath the slope cover, but if it does it has become much thinner. This geometry suggests that a gravel ridge existed before the deposition of the basal till. The origin of the ridge is unclear. Materials exposed there do not appear to be deformed as they are higher in the pit. In part this may be due to the coarse texture of the material. This lower ridge may have formed out in front of the advancing ice lobe due to ice shove or it may have been part of a pressure ridge formed by the weight of the ice lobe. Is it just a coincidence that the moraine parallel ridge lies above this deeper ridge?

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RIDGE, J. C., 1983, The surficial geology of the Great Valley section of the Valley and Ridge Province in eastern Northampton Co., Pennsylvania and Warren Co., New Jersey: unpublished M.S. thesis, Lehigh Univ., 234 p.

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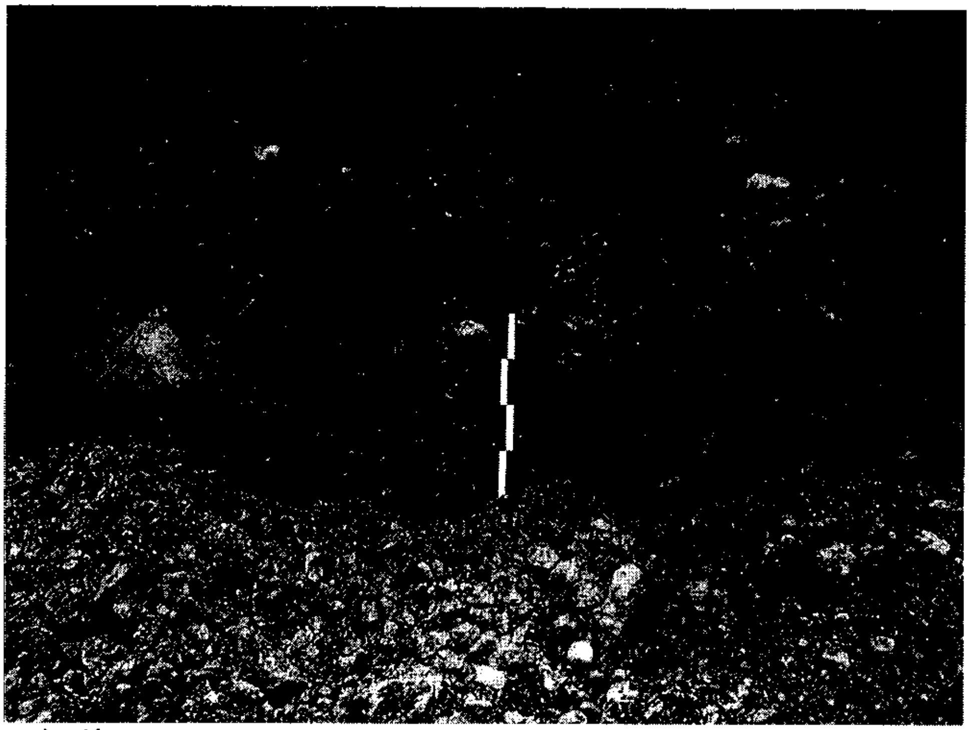

scale = 4 feet

Rose plot of the azimuth of elongated clasts (> 3 in.), measured in the upper till that makes up the Foul Rift moraine.

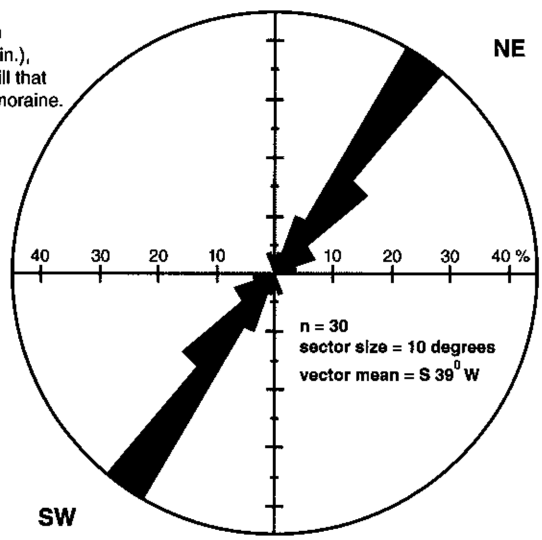

Figure VII-9. Upper till exposed along the north rim of the Foul Rift pit. Compact, fissile, sandy-silty matrix containing by volume 7 to 10 percent subangular to subrounded stones. Many stones are striated and elongated clasts have a pronounced downvalley fabric. Although the till forms part of the Foul Rift moraine, it has characteristics of a basal till. It may represent a subglacial till facies associated with a push moraine.

RIDGE, J. C., EVENSON, E. B., and SEVON, W. D., 1990, A model of late Quaternary landscape development in the Delaware Valley, New Jersey and Pennsylvania: Geomorphology, v. 4, p. 319-345.

SALISBURY, R. D., 1902, Glacial geology: New Jersey Geol. Survey, Final Report of the State Geologist, v. 5, Trenton, N.J., 802 p.

WARD, Freeman, 1938, Recent geological history of the Delaware Valley below the Water Gap. Pa. Geol. Surv., General Geology Report, 10, 65 p.

WITTE, R. W., and STANFORD, S. D., 1995, Environmental geology of Warren County, New Jersey: Surficial geology and earth material resources, New Jersey Geological Survey Open-file Map OFM 15C, 3 plates, map scale 1:48,000.

Retrace route back to the scale house or follow secondary route to north rim of pit where fresh exposures reveal the character of some morainal materials .

Continuation of road log.

27.8 Turn left, back onto Foul Rift Road and retrace route to Pequest Fish Hatchery.

28.5 Stop sign. Turn left onto County Route 519 North.

29.1 Intersection of County Routes 620 and 519. Turn right and continue on Route 519 North. Good view of Jenny Jump Mountain (looking about 10 o'clock on the horizon) on a clear day.

30.7 Stop light. Intersection County Route 623 (Brass Castle Road) and County Route 519. Continue straight on Rt. 519 north.

31.8 Turn left into Unagnst sand and gravel pit (just past township garage). Location of stop 5. Park buses in area near Route 519. Head back into upper pit along pit road. We do not have permission to access the lower pits.

STOP 5

Leaders: Scott Stanford and Ron Witte

Setting: The Pophandusing delta (fig. VII-10)sits on the east side of the Delaware Valley at the base of the New Jersey Highlands. The delta and its history were originally described by Ridge (1983). Based on its location the delta is coeval or slightly younger than the Terminal Moraine. Similar to the Terminal Moraine the delta was laid down from a retreat position of the Delaware Valley sublobe. The late Wisconsinan border lies about 1200 feet southeast of the delta on the lower slope of the Highlands and about 100 feet higher. Following ice retreat to the position marked by the delta or following a readvance (similar to that observed at the Foul Rift pit at stop 4), a small proglacial lake formed between the eastern edge of the Delaware Valley sublobe and the northwest facing slope of the Highlands. Based on its collapsed morphology the valley-side slopes of the delta were laid down against ice. Apparently the location of the delta represents a reentrant formed between ice in the main part of the Delaware Valley and a smaller sublobe that extended up the Pequest Valley and deposited the Bridgeville segment of the Terminal Moraine. In this ice-walled proglacial lake, meltwater, probably exiting the glacier through a tunnel, filled in the small lake basin with sand and gravel. Based on the geometry of the glacier's margin required to hold in the lake, and the elevation of its delta plain (485 feet), the level of the small lake was controlled by an outlet on a drainage divide between Buckhorn Creek and Pophandusing Brook. The delta plain in several places is marked by kettles, former blocks of stagnant ice that became detached from the main body of the retreating ice sheet. Further proof that the glacier did extend southward from the position shown by the ice-contact face of the delta.

Retreat from the Pophandusing delta position resulted in abandonment of the delta and a shift of meltwater deposition northward. A small ice-contact delta, elevation about 460 feet, laid down behind the Bridgeville segment of the Terminal Moraine marks this change. A meltwater channel cut through the eastern part of the Pophandusing delta, now utilized by Route 519, probably served as the outlet for this lower lake. Interbedded tills and ice-shoved foreset beds observed in the northern part of the Pophandusing delta by the authors (lower pits) show that margin of the Delaware Valley sublobe did oscillate throughout a narrow zone.

Retreat from the Bridgeville delta position resulted in the abandonment of the Pophandusing spillway and draining of the small proglacial lake in the Bridgeville area. Lower outwash deposits were laid from ice retreat positions located farther northward in the Beaver Brook and Delaware River valleys and from meltwater (largely from Lake Pequest) coming down the Pequest Valley through Pequest Gap.

Materials: The sediments exposed in the upper (Unangst) pit are fine-to-coarse sand with some granule gravel and a little pebble-to-fine cobble gravel, and rare silt and clay laminae. The sand and gravel are arranged in a stacked, cross-cutting series of lobes and swale fills, with the bases of many of the swales channeled into underlying material (fig. VII-11). Bedding within these sets of lobes and swales is chiefly large-scale foresets, planar lamination, or curved swale-fill bedding fitted to the channeled base of the swale. Ripple cross-bedding is absent, although in a few beds the planar lamination appears to be disrupted or deformed into a wavy lamination. In one or two spots, silt and clay laminae drape the lobe and swale bounding surfaces. Gravel typically occurs in thin beds at or near the base of the swales, on the channeled surfaces. The gravelly units are weakly graded, and tend to fine upward to sands. In a couple of spots there are thin diamictons interbedded with the sands. At the very top of the pit is a thin (< 4 feet thick), massive medium-to-coarse cobble gravel. At the north end of the pit there is a thin reddish yellow silty till that veneers the landsurface and lies, with sharp lower contact, on top of the stratified sediments.

The lower pit, which we will not visit, has been worked for many years. Exposures there have shown similar lobe-and-swale bedding as in the upper pit, but have also exposed beds of coarse, openwork (i.e., matrix-poor) cobble gravel as much as 10 feet thick, with channeled lower contact, within the sandy lobe-and-swale deposits. The overlying till has also been well-exposed, again in sharp contact with the underlying stratified deposits, although it rarely is more than 3 feet thick.

Gravel in both pits is dominantly gray carbonate rock, gray to brown shale and sandstone, and gray, white, and red quartzite and conglomerate. This composition indicates a Delaware Valley source for the meltwater feeding this deposit, with little contribution from the gneiss uplands to the east.

Discussion: The bed sets and bedforms observed in these pits are characteristic of lacustrine fans. These fans are built at the mouths of subglacial tunnel channels that discharge beneath lake level at the glacier margin. Because these tunnel channels are discharging beneath lake level, they require hydraulic head higher than the lake level to initiate flow. This head is built up within the glacier by meltwater draining down into, and filling, crevasses and

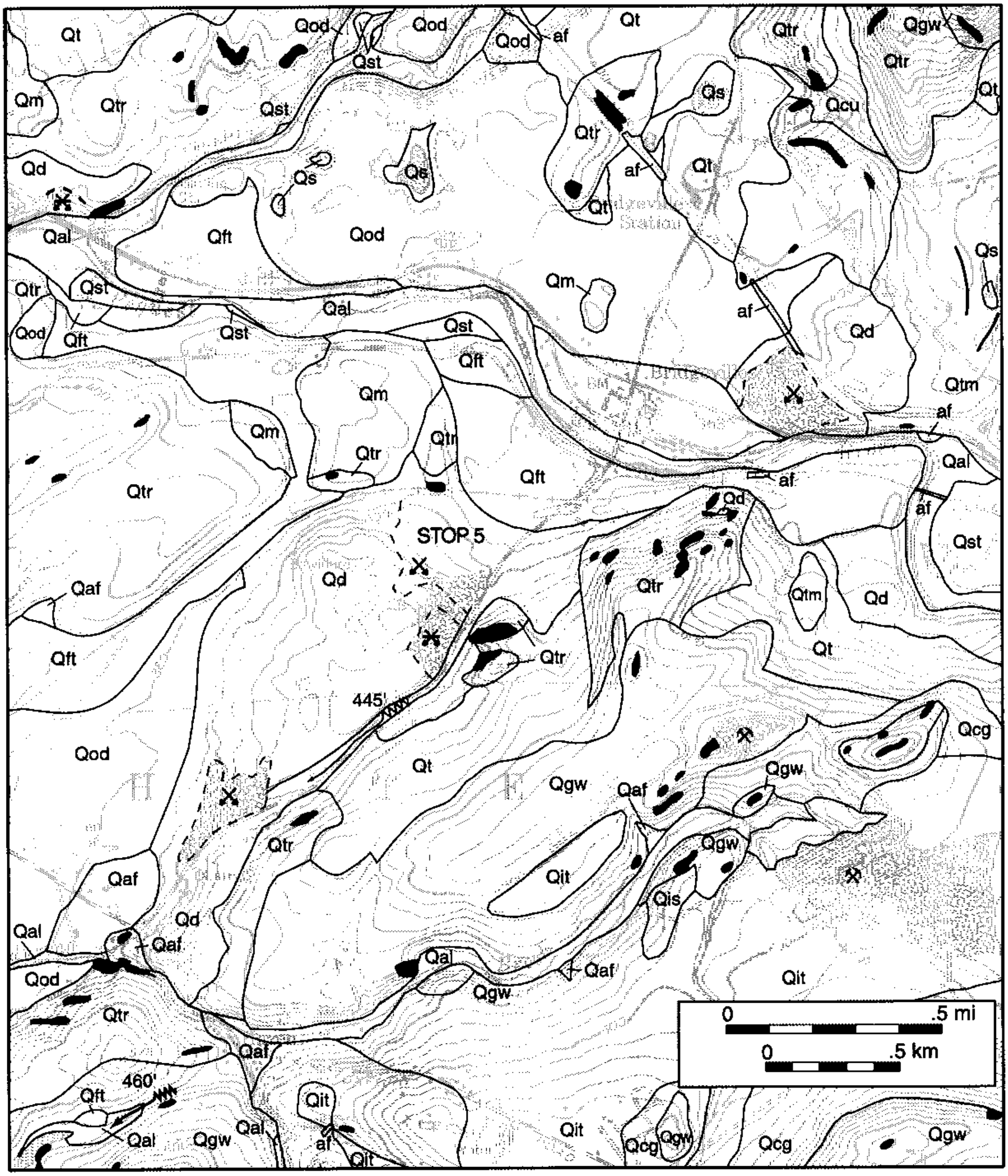

Figure VII-10. Surficial geologic map of part of the Belvidere, NJ-Pa, 7 1/2 minute topographic quadrangle near Bridgeville, New Jersey and location of stop 5. List of map units: af - artificial fill, Qal - alluvium, Qaf - alluvial fan, Qst - stream-terrace deposit, Qs - swamp deposit, Qtm - Terminal Moraine, Qt - thick till, Qtr - thin till, Qd - ice-contact delta, Qod - valley-fill delta, Qft - meltwater-terrace deposit, Qit - till of Illinoian age, Qcg - gneissic and granitic colluvium, and Qgw - weathered gneiss and granite. Areas shaded black represent bedrock outcrops. Thin lines with an arrow denote lake spillways. Data modified from Witte and Ridge (in preparation), Surficial geologic map of the Belvidere Quadrangle, NJ - PA.

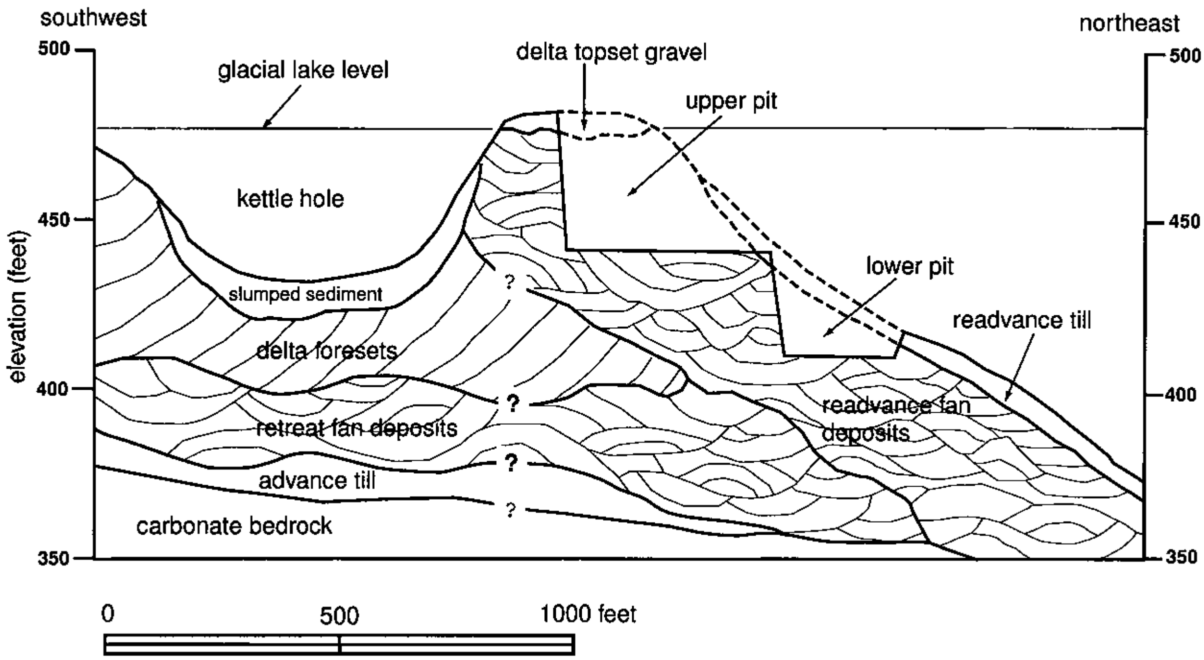

Figure VI-6.--Cross section of fan-delta deposit at stop 5. Thin detail lines show, in diagrammatic fashion, bedding in the fan and delta sediments. Queried contacts, and the units they separate, are hypothesized based on local geologic relations. Delta foresets prograde southward from the readvance fan, across earlier fan deposits laid down during retreat from the late Wisconsin maximum position. The readvance fan is built up over the retreat fan deposits as ice readvances to the south. This readvance is marked by a thin till at the surface, in erosional contact with the readvance fan deposits. The fan deposits are characterized by cross-cutting sets of channels and lobes that record the shifting position and sharply varying discharge of the tunnel channels.

fractures. The meltwater collects at the glacier bed and discharges to the margin through subglacial channels. Ice, and wet basal sediment, flow toward the tunnel channel but the tunnel is kept open by the water flow. The inflowing ice and sediment deliver material to the channel. Also, because the flow is occurring below the piezometric surface, it is closed-channel, pipelike flow, not open-channel stream flow.

These hydraulic conditions create distinctive bedforms (Gustavson and Boothroyd, 1987; Ashley and Shaw, 1988). Because flow will not occur below a threshold head, tunnel channels experience rapid alternations between high-velocity flow (when head exceeds the initiating threshold) and no-flow (when head drops below the threshold). For this reason, most of the bedforms are high-velocity planar beds, with little or no ripple cross-bedding. Because the tunnel is discharging into quiet lake water, velocity drops rapidly at the tunnel mouth and sediment accumulates, clogging the discharge point. This causes frequent current shifts, accounting for the cross-cutting, channeling, and stacking of lobe and swale bed sets. The large-scale foreset beds that dominate some of the sets probably represent progradation of the lobes out into the lake, away from the discharge point. The general southerly and southwesterly progradation direction of the foresets in the upper pit is consistent with tunnel channels discharging out to the glacier margin from the Delaware Valley to the north and northwest (fig. VII-1). The thick, coarse cobble gravel beds in the lower pit may have been deposited within the tunnel channel itself. Similar gravels have been described from subglacial esker deposits (Ashley and others, 1991; Warren and

Ashley, 1994).

The local geologic relations in this area indicate that the exposed fan deposits were laid down in front of a minor readvance of the ice front, following retreat from the southernmost limit. This readvance, which is marked by the thin till lodged on the north slope of the fan, may be the same as the one marked by the upper till at stop 4. The readvance fan built up to lake level, as indicated by the presence of topset gravels at the surface of the upper pit (fig. VII-6). Once the lake level was reached, sediment delivered by the tunnel-channel system prograded into the lake as sandy delta foresets. Foreset sand is exposed in pits farther south in the deposit.

Lacustrine-fan sediments are important hydrogeologic units in New Jersey (see Chapter IV). In most lake basins, lacustrine fans are laid down as linear, beaded bodies that track the retreating tunnel mouth. They generally occur on the rock or till floor of the lake, often in the lowest part of the lake basin. For this reason, they are commonly covered by lake-bottom deposits or by prograding delta deposits. These lower-permeability clay-silt, or silt-fine sand, sediments act as confining or semi-confining layers. Thus, the lacustrine-fan deposits may be productive confined aquifers, particularly if a thick, openwork, cobble-gravel channel deposit is tapped.

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Gustavson, T. C., and Boothroyd, J. C., 1987, A depositional model for outwash, sediment sources, and hydrological characteristics, Malaspina Glacier, Alaska: a modern analog of the southeastern margin of the Laurentide ice sheet: Geological Society of America Bulletin, v. 99, p. 187-200.

Warren, W. F., and Ashley, G. M., 1994, Origins of the ice-contact stratified ridges (eskers) of Ireland: Journal of Sedimentary Research, v. A64, no. 3, p. 433-439.

Continuation of road log.

31.8 Turn left onto County Route 519 North from parking area.

32.3 Cross Pequest River.

32.4 Stop light. Intersection of County Route 519 and U.S. Route 46. Turn right onto Route 46 East.
33.7 Stop light. Intersection of U.S. Route 46 and State Route 31. Continue straight on Route 46 East.

36.4 Pass entrance to Pequest Fish Hatchery on the south (right) side of U.S. Route 46.

36.6 Cross over late Wisconsinan border.

36.7 Pass a large outcrop of dolostone (Allentown Formation) on north (left) side of the highway. The low hill above the outcrop make up part of an ice-contact delta laid down in Lake Oxford. The delta marks the late Wisconsinan border in Pequest Valley.

38.2 Good view of Pequest River and its flood plain on the southeast (right) side of the highway. The steep scarp on the far side of the river has been cut in the Townsbury segment of the Terminal Moraine. The Pequest River here probably follows a meltwater channel cut by outlet waters from Lake Pequest.

38.7 J.P. Trucking company on southeast (right) side of the highway. 1984 Friends of the Pleistocene field stop. In the sand and gravel pit on the property, several deltaic/lacustrine facies were observed. These included coarse gravelly fan foreset beds, bottomset beds of laminated, silty fine sands with drop stones, and lower foreset beds of fine to coarse sand. These deposits represent a coarsening upwards sequence laid down in Lake Pequest (stage I).

39.1 Pass entrance to Pia Costa sand and gravel pits on right. These pits are in the Pequest Union Cemetery delta. The rolling hills on the southeast (right) side of the highway are underlain by sand and gravel. The topography here is the result of collapse where deltaic sediment was formerly laid down over stagnant ice. U.S. Route 46 here follows a meltwater channel cut in the delta. The channel served as a sluiceway for Lake Pequest (stage II).

39.3 Enter the village of Great Meadows.

39.4 Intersection of County Route 611 and U.S. Route 46. Turn left onto Route 611(Hope Road). Danville Mountain lies to the south (left) and floor of Lake Pequest to the north (right). Thin deposits of peat (alluvial origin) and sandy alluvium overlie lake-bottom deposits (glacial lake varves, as much as 200 feet thick). The valley floor has been drained by a series of interconnected ditches for flood control and agricultural use(chiefly sod and vegetables).

40.9 Pass intersection of County Route 611 and Shades of Death Road on north (right) side of the road.

41.4 Pass intersection of County Route 611 and Marble Hill Road on south (left) side. Cross over Marble Hill. The topography here is chiefly formed by thin till on bedrock.

41.4 Climb east flank of Jenny Jump Mountain.

41.7 Intersection of County Route 611 and Farview Road on north (right) side of road. Turn Right onto Farview Road, heading toward Jenny Jump State Forest. Note change in the character of landscape to that of bedrock controlled topography. There are many glacially scoured outcrops of gneiss.

42.8 Enter Jenny Jump State Forest

43.2 Turn right into entrance for Jenny Jump State Forest.

43.3 Visitor's Center. From here we will disembark the buses and continue to the beginning of Summit Trail where we'll reassemble for a short discussion on the Park's history. Afterwards we'll go on to the overlook via Summit Trail where we'll discuss glacial erosion. Please use Figure VII-12 in the guide-book to locate several glacial erosional features along the trail.

STOP 6
Jenny Jump State Forest: Summit trail and overlook
Glacial erosional features

Leaders - Ron Witte, and Patty Bennett (N.J. Div. of Parks and Forestry)

Setting: Jenny Jump State Forest (fig. VII-1) consists largely of wooded acres atop Jenny Jump Mountain, Warren County, New Jersey. The mountain makes up part of the Highlands physiographic province; a rugged upland chiefly underlain by metasedimentary and intrusive rocks of Proterozoic age. Jenny Jump Mountain, Danville Mountain, High Rocks Mountain, Mount Mohepinoke, and several smaller, unnamed hills collectively form the Jenny Jump Thrust sheet (Drake and others, 1996). During the Alleghenian orogeny, crystalline rocks were thrust up and over sedimentary strata of Lower Paleozoic age. Because the gneiss and granite that hold up the mountain are more resistant to erosion than the surrounding sedimentary strata (dolostone and shale) it stands out in bold relief rising as much as 700 feet above adjacent valley floors.

Locally, Jenny Jump Mountain consists of several narrow ridges that trend southwest. Topography is rugged due to the many outcrops of glacially-sculpted rock. Thin till deposited during the late Wisconsinan glaciation lies on the lower parts of rocky slopes and hollows between rock ridges. On the mountain's northwest flank, thicker till forms an extensive blanket masking the rock-controlled topography. In several places, wind gaps as much as 150 feet deep, cut through the mountain. High above the modern valleys, these features are relics from a much older landscape.

Glacial History: During the last ice age Jenny Jump Mountain was covered by the Laurentide ice sheet, except a small area near its southwest termination (Ridge, 1983), just beyond the Jenny Jump segment of the Terminal Moraine. During the maximum extent of the ice sheet, ice at this location would have been 1000 to 1500 feet thick (see Chapter III for a more detailed discussion on ice sheet's surface gradient near its margin). While the mountain was ice covered, the ice sheet slowly eroded the preglacial soil cover, removed loosened bedrock, and sculpted the rock outcrops into streamlined forms. During deglaciation the margin of the Kittatinny Valley lobe retreated northeastward. In the Pequest Valley, Lake Pequest formed between the Townsbury segment of the Terminal Moraine and the Pequest sublobe. In places the lake was as much as 300 feet deep; its deposits now form the floor of the Pequest Valley north of and including Great Meadows.

Discussion: Throughout the previous five stops, we have looked at glacial depositional processes and constructional glacial landscapes within the framework of advance and retreat of the Laurentide ice sheet. Stratified glacial materials and their depositional environments are largely studied because they hold the greatest economic value (natural aggregate, prolific aquifers, level areas to build). However, the largest areas that make up the glacial landscape include areas of thin till and bedrock. Because the greatest work of an ice sheet is in its ability to erode soil, loose rock and bedrock and transport it elsewhere. At this stop we'll examine evidence of glacial erosion, specifically erosion of bedrock and discuss its large- and small- scale forms.

The competence of an ice sheet to erode preglacial soil and rock is easily grasped if one compares areas recently glaciated with those that were glaciated much longer ago. This setting is shown in Figure VII-13, which shows that the area covered by rock outcrops north of the late Wisconsinan border constitutes a much higher percentage of the land. Over time, the effects of weathering will transform this glacial landscape to an older colluvial landscape, where outcrops are sparse due to thickening regolith.

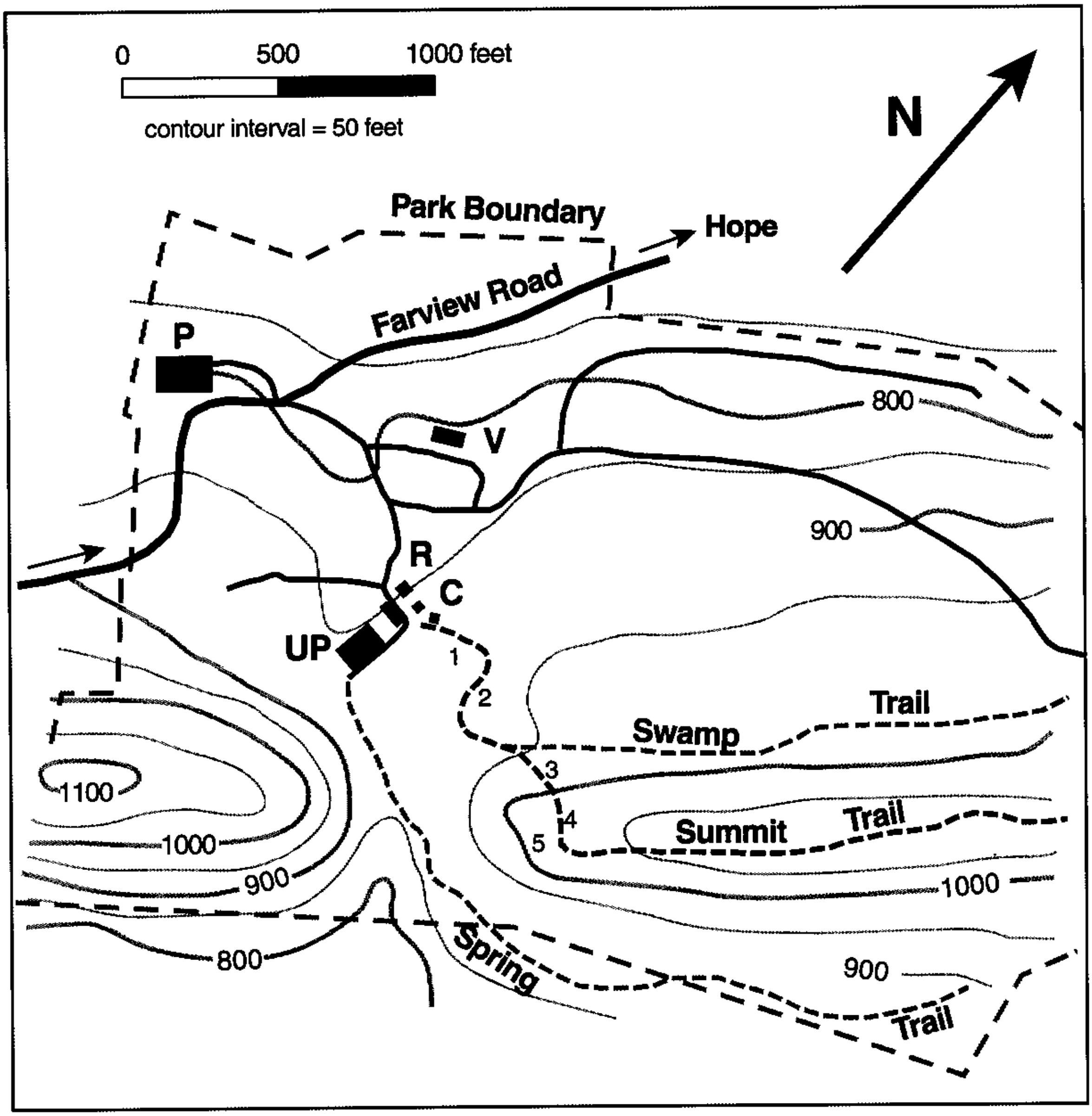

Figure VII-12. Location of STOP 6, Jenny Jump State Forest and route to overlook off Summit Trail. Numbered areas along trail denote features described in guidebook, Chapter VII, stop 6. Key to symbols: V - Visitor's Center, P - Parking area for buses after drop off, UP - upper parking area, R - rest rooms, and C - cabins near Summit Trail.

Glacial erosion of bedrock occurs chiefly by abrasion (the mechanical wearing of rock by grinding and scraping by rock particles transported at the base of an ice sheet) and plucking (the process of glacial erosion by which blocks, and large rock particles are loosened, detached, and carried away from bedrock). Under most conditions abrasion occurs on the up-ice (stoss) side of the outcrop, and plucking occurs on its down-ice (lee) side. These forms of glacial erosion give the glacial-sculpted rock its characteristic whaleback shape. The rock ridge and swale topography formed over the mountain was largely formed by the glacial erosion of coarse-grained gneiss and granite that have an abundance of subvertical joints that bisect the rock's layering. Pre-glacial topography and lithologic variation within formations and between formations also contributed to the formation of this glacial erosional landscape.

We estimate that Jenny Jump Mountain at this location may have been covered by ice for as long as 4000 years. Rates of erosion have been estimated to range between .06 to 5 mm per year (Ritter, 1978, p. 378). Using the higher number, Jenny Jump Mountain may have been lowered by as much as 2 meters during the last glaciation. This amount does not take into account the removal of regolith, only non-weathered bedrock. Because the effects of glacial climates act on rock to break it up (largely by cyclic freeze and thaw) and prepare it so that it can be more easily eroded, the rate of erosion may be about a magnitude higher (20 m).

Glacial erosional features found along Summit Trail (Figure VII-12) include:

1 - Start of Summit Trail. Glacially sculpted outcrops of gneiss (fig. VII-14) and joint-block boulders (fig. VII-15) on the right side of trail.

2 - Glacially-polished gneiss bedrock pavement on trail. Glacial grooves here show that ice flowed south 30 to 35 degrees west.

3 - Small glacially-sculpted gneiss outcrop. Note whaleback form.

4 - Large polished gneiss pavement (fig. VII-16a) at intersection of Summit Trail and trail to overlook. Faint grooves on upper side of pavement (fig. VII-16b) show that ice flowed South 34 degrees West. Bedrock foliation here trends south 50 to 55 degrees west.

Other features that may be seen along trail include:

- glacial erratics (rocks carried by glacial ice, and deposited at some distance from the outcrop from which it was derived). Includes rocks from Kittatinny Valley (dolostone, chert, and sandstone), and rocks from Kittatinny Mountain (quartzite, quartz-pebble conglomerate, and red sandstone). A few dolostone erratics in the park are more than 20 feet in diameter.

5 - The overlook (location 5 on fig. VII-12) affords an exceptional view of the Pequest Valley, and some mountains that make up the New Jersey Highlands. In the valley, the broad flat land to our left forms the Great Meadows area, the southern part of Lake Pequest. Low hills of dolostone form islands across the meadows. Collectively they define a rock ridge that trends northeastward into the upper part of the lake basin. The wind gap beneath the overlook is about 150 deep and it cuts across the mountain's northeastward topographic grain. The gap's floor lies 850 feet above sea level, about 350 higher than the floor of the Pequest Valley. If the gap was originally cut by a stream then it is a relic of a much earlier pre-Pleistocene time (see Chapter III for details on the preglacial drainage). The gap's location is also coincident with the primary joints of the local bedrock. Perhaps the frequency of jointing in this part of the ridge was such that a stream could downcut more easily and take advantage of the weaker rock.

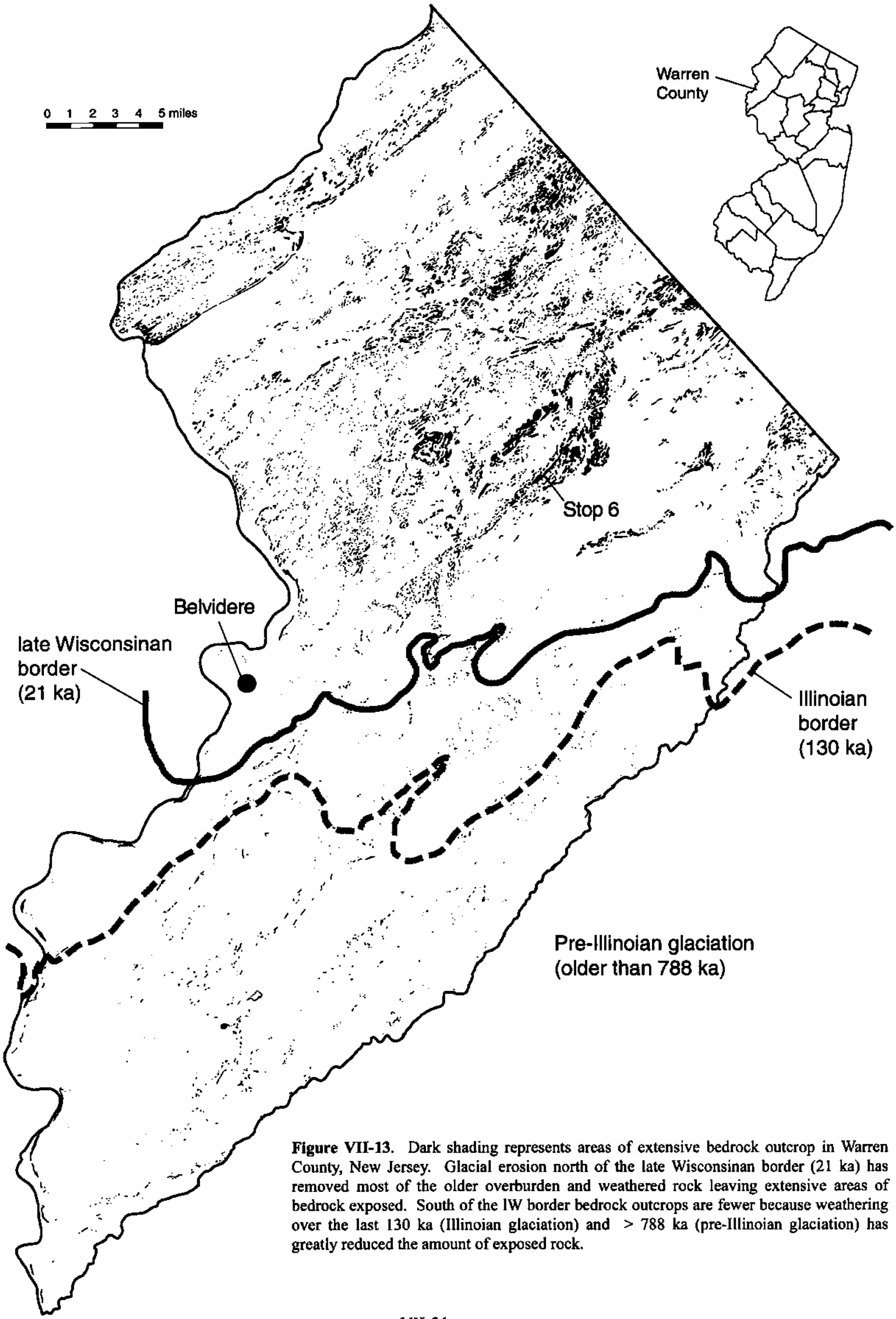

Figure VII-13. Dark shading represents areas of extensive bedrock outcrop in Warren County, New Jersey. Glacial erosion north of the late Wisconsinan border (21 ka) has removed most of the older overburden and weathered rock leaving extensive areas of bedrock exposed. South of the lW border bedrock outcrops are fewer because weathering over the last 130 ka (Illinoian glaciation) and > 788 ka (pre-Illinoian glaciation) has greatly reduced the amount of exposed rock.

Figure VII-14. Glacially-sculpted gneiss outcrop located near the start of Summit Trail (location 1 on Figure VII-12). Whaleback shape of outcrop is marked by heavy black line. Scale (upper right of photo equals two feet). This charateristic erosional shape is produced by glacial abrasion (sides and up-ice side of the outcrop) and glacial plucking (steep face on the down-ice side of outcrop). Nearly vertical jointing that trends oblique to the direction of ice flow (see figure VII-15 below) helps accentuate plucking.

Figure VII-15. Joint-block boulders formed on the down-ice-side of glaccially sculpted bedrock ridge. Photograph taken just to the right of Figure VII-14. Scale in center equals two feet. Through successive cycles of freeze and thaw, ice expanded horizontal and vertical fractures. Eventually, blocks of stone were detached from the main body of bedrock, some of which now lie at the base of the outcrop where they have fallen. Rock to the left of the scale is in place.

Figure VII-16a. Polished gneiss pavement at the intersection of Summit Trail and trail to overlook (location 4 on Figure VII-12). Glacial polish consists of numerous microscopic scratches formed by silt and sand at the glacier's base sliding over bedrock. Postglacial weathering has roughened this surface slightly, although the footsteps of many hiker's have also polished the rock.

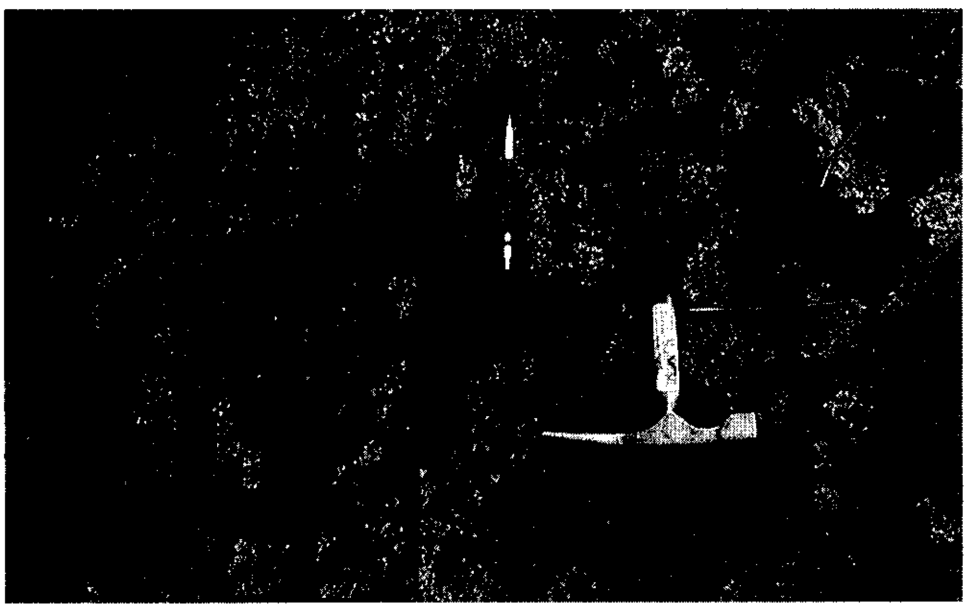

Figure VII-16b. Glacial grooves cut in gneiss pavemnt (location of photograph is shown in the above figure). Gooves and scratches form where larger rock particles, entrained in ice at the glacier's base, cut linear furrows in the underlying rock. These stones must be as hard or harder than the rock they are overriding. The grooves here indicate ice flowed south 34 degrees west when they were cut; a direction that cuts across the more southwesterly topographic grain of Jenny Jump Mounatin.

The overlook area is formed on a large glacially-sculpted (whaleback) bedrock ridge. We are standing on the plucked or down-ice side of the ridge. The local bedrock consists of pyroxene gneiss (Drake and others, 1996). It is a well-layered metasediment, containing oligoclase, clinopyroxene, and variable amounts of quartz. Some phases of the rock contain calcite. Foliation is nearly vertical and trends North 55 to 70 degrees East and primary joints are also nearly vertical and they trend about North 56 degrees West. Postglacial weathering has slightly-roughened the glacially-polished gneiss surface, obliterating most of the striations. In places, a thin soil derived from the fragmental disintegration of the coarse-grained gneiss, fills in some fractures and lies below some plucked-rock faces. Mechanical weathering, chiefly through successive cycles of freeze and thaw, has dislodged some joint blocks. In many parts of the park these boulders form aprons of rock rubble around the outcrop

Glacial erosional features and other geologic oddities found at the overlook.

- lunate gouge (fig. VII-17)

- differential erosion on gneiss surface (fig. VII-18).

- postglacial weathering of scour depressions (fig. VII-19).

- horizontal fractures possibly formed by glacial unloading (fig. VII-20)

- crescentic fractures - type of chattermark that consists of a convex-shaped single fracture in the rocks surface. Horns point in the direction that ice moved

References Cited

DRAKE, A. A., Jr., VOLKERT, R. A., MONTEVERDE, D. H., HERMAN, G. C., HOUGHTON, H. H., PARKER, R. A., and DALTON, R. F., 1996, Bedrock Geologic Map of Northern New Jersey: U.S. Geological Survey Misc. Geol. Inv.Map I-2540-A.

RITTER, DALE, F., 1978, *Process Gemorphology*, Wm. C. Brown Compay Publishers, Dubuque, Iowa, 603 pp.

Continuation of road log.

43.5 Turn right onto County Route 611 from entrance to Jenny Jump State Forest. Descend northwest flank of Jenny Jump Mountain. In places, thick till forms an extensive cover, masking the bedrock topography on the mountain's northwest-facing slope

44.4 Cross onto carbonate bedrock. Note change to a karst landscape, characterized by sinkholes, solution valleys, and pinnacles of dolostone.

44.5 Stop sign. Turn left onto Shiloh Road.

45.7 Stop sign. Turn left onto County Route 519 South.

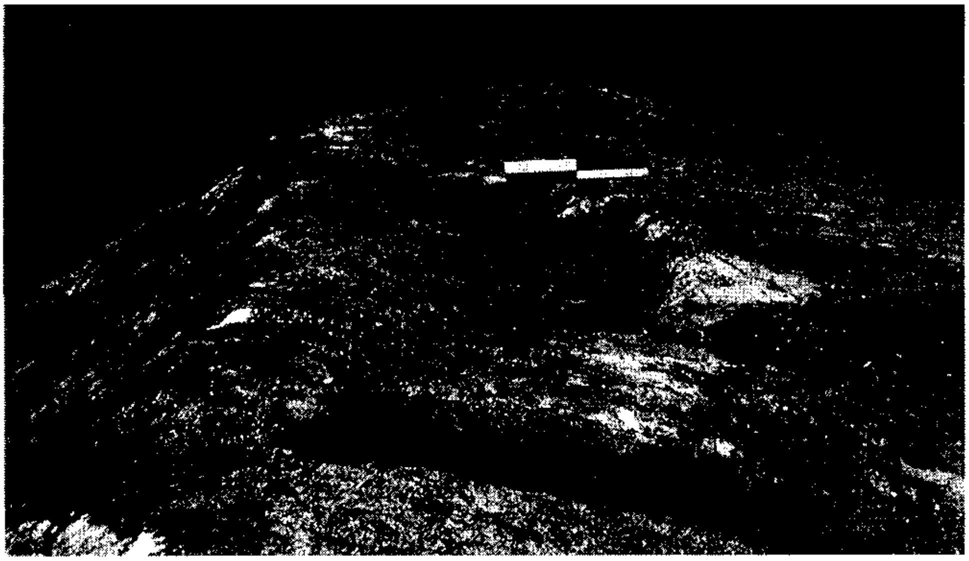

Figure VII-17. Lunate gouge formed on gneiss outcrop at overlook off Summit Trail (location 5 on Figure V11-12). Scale equals two feet. Ice flow was from right to left. These features are large chips, possibly formed where boulders at the base of and under the extreme weight of the the ice sheet, may have been in contact with the gneiss.

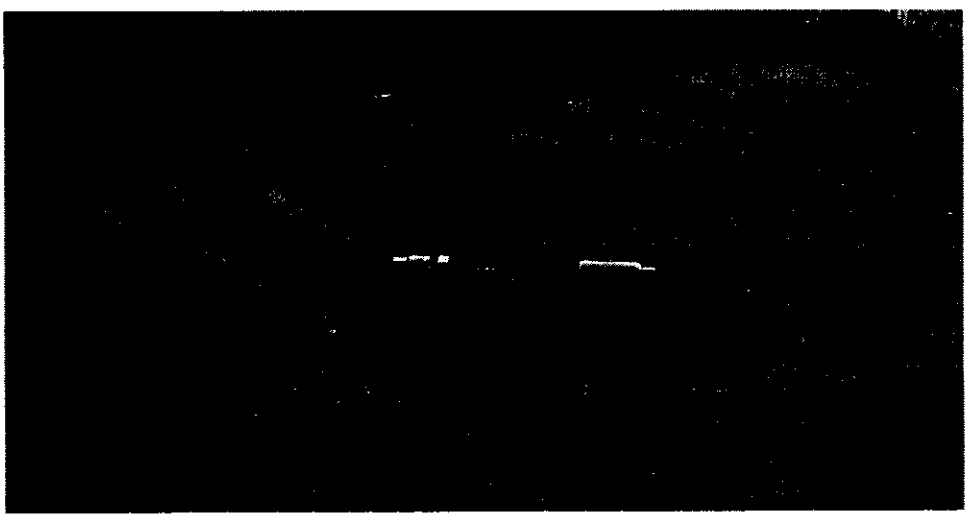

Figure VII-18. Quartz-rich foliation bands stand out as much as 0.5 inches on gneiss surface (location 5 at overlook off Summit Trail, Figure VII-12). Most of the relief seen here is due to glacial abrasion Because the quartz is harder than the surrounding mix of minerals, it forms higher areas on the outcrop's surface. Ice flow was left to right.

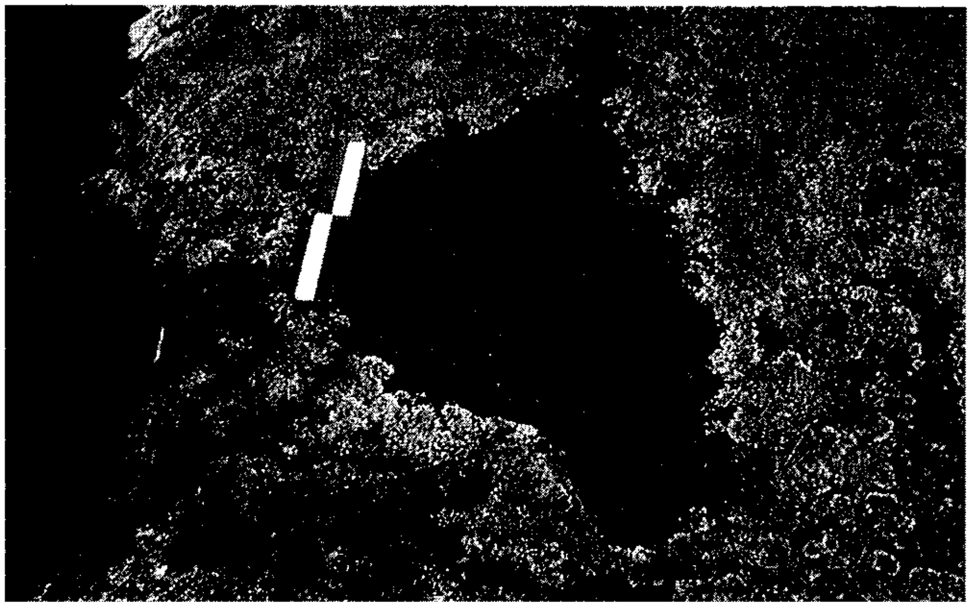

Figure VII-19. Minor enlargement of *sichelwannen* by postglacial weathering. Rain water periodically collects in the shallow depression and throughout many cycles of wetting and drying some of the crystalline rock is removed by dissolution.

Figure VII-20. Subhorizontal fractures exposed at the plucked face of a glacially-sculpted gneiss ridge. Scale equals two feet. These fractures may have formed during isostatic unloading and crustal rebound, related to the melting of the Laurentide ice sheet.

46.4 Enter the village of Hope, New Jersey.

46.7 Stop light, intersection County Routes 519 and 521. Turn right onto Route 521 North.

47.7 Entrance on to ramp for Interstate Route 80 east.

End of road log.